MAP REPORT 81-3

QUATERNARY GEOLOGY OF AMHERST MAP AREA (21 H), NEW BRUNSWICK

V. N. RAMPTON

S. PARADIS

(TERRAIN ANALYSIS AND MAPPING SERVICES LTD.
CARP, ONTARIO)

Government of Canada / Gouvernement du Canada
Regional Economic Expansion / Expansion Économique Régionale

NATURAL RESOURCES
NEW BRUNSWICK
RESSOURCES NATURELLES
NOUVEAU-BRUNSWICK

1981

This is one of a series of map reports produced by the New Brunswick Department of Natural Resources and is jointly funded by the Department of Regional Economic Expansion (DREE), Government of Canada, and the Department of Natural Resources, Government of New Brunswick.

Mineral Development Branch
Department of Natural Resources
Fredericton
New Brunswick
1981

CONTENTS

	Page
Abstract/Résumé	v
Introduction	1
Acknowledgements	1
Previous Work	1
Physiography, Bedrock and Drainage	2
Surficial Geology	4
Description of Map-Units	4
Bedrock (R)	5
Undifferentiated Deposits and Landforms	6
Colluvium (C)	6
Glacial Deposits and Landforms	6
Morainal Deposits (M)	6
Lodgement Till	7
Ablation Moraine	7
Glaciofluvial Deposits (G)	8
Ice-Contact Deposits	8
Outwash	9
Late Glacial and Postglacial Deposits and Landforms	9
Submerged and Wave-Washed Surfaces	9
Marine and Lacustrine Deposits (W,L)	10
Postglacial Deposits and Landforms	11
Organic Deposits (O)	11
Alluvial Deposits (A)	11
Marine Deposits (W)	12
Interpretation of Quaternary History	12
Introduction	12
Early Sea Levels	13
Early Glaciation	13
Wisconsin Glaciations	13
General Chronology	13
Nomenclature	13
Sources of Glaciers in Southeastern New Brunswick	15
Caledonia Phase	15
Chignecto Phase	20
Kent Phase	22
Bantalor Phase	26
Sackville Phase	30
Late Glacial and Postglacial Events	30
Late Wisconsin and Early Postglacial Submergence	30
Postglacial Emergence	33
Late Postglacial Submergence	33
Terrestrial and Alluvial Events	33
References	34

Tables

Table 1.	Radiocarbon Dates and Descriptions	23

Illustrations

Map 81-003	Quaternary Geology, Amherst	in pocket
Figure 1:	Physiography of southeastern New Brunswick	3
2:	Intertidal sediments along Tantramar River	14
3:	Wave-planated rock bench southwest of Rockport	14
4:	Drainage patterns near South Branch	16
5:	Drainage patterns near Berryton	16

Page

Figure 6: Glacial features of Caledonia phase 17
7: Chignecto glacial limit near Weldon Creek 19
8: Glacial features of Chignecto phase 21
9: St. Martins and Pollett River flow patterns and glacial features of Kent phase 25
10: Glacial features of Bantalor phase 27
11: Glacial features of Sackville phase and ice marginal features of Bantalor retreatal phases 29
12: Indicators of marine submergence 31

ABSTRACT

Data acquired during the mapping of the Quaternary geology of the Amherst map area provides a basis for the Quaternary history of southeastern New Brunswick. A marine bench that predates the earliest recorded glaciation has been identified 4 m above its modern counterpart. Subsequently, during the Caledonia phase of glaciation, the Caledonian Highlands were glaciated by southeast flowing ice. The periphery of the Caledonian central plateau was then deeply incised by streams prior to further glaciation. Afterwards, during the late or middle Wisconsin, glaciers flowed southwest into Chignecto Bay from an ice centre near Point Esecuminac, and south near St. Martins during the Chignecto and Kent (?) phases. In late Wisconsin during the Bantalor phase, glaciers from the northwest and from the Gaspereau ice centre in south central New Brunswick dominated. These glaciers began retreat from their maximum position near Hammond River around 13,500 B.P. During ice retreat a number of major pauses occurred and ablation moraines and ice-contact deposits were deposited along the ice margins. Ice also readvanced after 13,500 B.P. from an ice centre in Northumberland Strait to near Sackville. Due to glacio-isostatic perturbations the area was initially submerged after deglaciation and then emerged as rebound proceeded. Submergence has been occurring over the past 4000 years.

RÉSUMÉ

Les données recueillies au cours des levées cartographiques dans le sud-est du Nouveau-Brunswick sont à la base de la reconstitution de l'histoire du Quaternaire de la carte Amherst (21 H). On a identifié une plate-forme d'abrasion marine, a 4 m au-dessus de celle d'aujourd'hui et antécédente à la plus ancienne glaciation reconnue. Par la suite, lors de la phase glaciaire Calédonia, les hautes-terres calédoniennes furent envahies par des glaces s'écoulant vers le sud-est. Suivit une période pendant laquelle le pourtour du plateau central calédonien fut profondément entaillé par des cours d'eau. Plus tard, au milieu ou vers la fin du Wisconsin, lors des phases Chignecto et Kent (?) et à partir de la calotte situé près de la Pointe Escuminac, des glaciers s'écoulèrent en direction sud-est dans la baie de Chignecto et franc sud vers St. Martins. A la fin du Wisconsin, les glaciers en provenance du nord-ouest et de la calotte Gaspereau situé au centre-sud du Nouveau-Brunswick dominèrent. Ces glaciers entamèrent leur retraite a partir de leur position maximale près de Hammond River, autour de 13,500 B.P. De nombreuses pauses accompagnèrent ce retrait, entrainant la mise en place de moraines d'ablation et de dépôts juxta-glaciaires en bordure du front glaciaire. Peu après 13,500 B.P., une réavancée s'effectuait à partir du centre glaciaire situé dans le détroit de Northumberland jusqu'aux environs de Sackville. Des perturbations glacio-isostatiques affectèrent la région qui fut initialement submergée après la déglaciation pour ensuite émergé grâce au relèvement isostatique. La submersion à débuté il y a environ 4000 ans.

QUATERNARY GEOLOGY OF AMHERST MAP AREA (21 H), NEW BRUNSWICK

INTRODUCTION

In June of 1979, Terrain Analysis and Mapping Services Limited was instructed to proceed with the surficial mapping of central and southeastern New Brunswick, namely the New Brunswick portions of Amherst (21 H), Moncton (21 I) and Woodstock (21 J) map areas. This was accomplished by aerial photograph interpretation in the summer of 1979 and spring of 1980, a preliminary field investigation in the fall of 1979, extensive field work during the spring, summer and fall of 1980, review of earlier studies on the subject, and production of final maps and reports during late 1980 and early 1981. Aerial photograph interpretation and map compilation was completed by S. Paradis and V.N. Rampton; field work by S. Paradis, D.P. Smith and V.N. Rampton; radiocarbon dating by Brock University, St. Catherines, Ontario and Saskatchewan Research Council, Saskatoon, Saskatchewan; cartography by C. Ulmann and Maritime Resource Management Service, Amherst, Nova Scotia; report writing by V.N. Rampton; editing by V.N. Rampton, J.M. Rampton and M. Corrigan; and report and map printing and reproduction by Beauregard Press Limited, Ottawa. Field assistance during 1980 was provided by Robert J. Dubasz and James J. Richard of Fredericton.

ACKNOWLEDGEMENTS

We would like to thank Donald E. Barnett (Mineral Development Branch, Department of Natural Resources) for his co-operation in supervising the project; Dr. R.J. Fulton (Geological Survey of Canada) and Donald E. Barnett, Jacques Thibault, Allen Seaman (Mineral Development Branch, Department of Natural Resources), and B.M.W. Carroll (Geological Survey Branch, Department of Natural Resources) for reviewing and editing the report; Jacques Thibault, Paul Finamore (formerly of Mineral Development Branch, Department of Natural Resources) and Claude Gauthier (Geological Survey of Canada) for showing and reviewing relevant locations within and adjacent to the project area with V.N. Rampton; Jacques Thibault, Paul Finamore, Claude Gauthier, Dr. N.R. Gadd (Geological Survey of Canada), Dr. V.K. Prest (formerly of Geological Survey of Canada), and Dr. D.R. Grant (Geological Survey of Canada) for the insight gained from discussions of the local and regional Quaternary geology; Jacques Thibault, Drs. N.R. Gadd, V.K. Prest and D.R. Grant for providing unpublished documents and publications not easily available from libraries; Drs. D.V. Venugopal and A.A. Ruitenburg (Geological Surveys Branch, Department of Natural Resources) for their opinions on probable sources of erratics collected in eastern New Brunswick; R.J. Mott (Geological Survey of Canada) for providing unpublished radiocarbon dates; Dr. Weston Blake, Jr. (Geological Survey of Canada) for examining and identifying shells, and providing radiocarbon date GSC-3167; Mrs. M.F.I. Smith (National Museum of Canada) for identifying and commenting on the range of the species whose shells were used to obtain GSC-3167; Howard Melville (Brock University) for courteous service and expedient provision of radiocarbon dates; Cartographic Division, Maritime Resource Management Service for their interest, direction and co-operation in preparing negatives for the colour map; Beauregard Press Limited for their co-operation in reproduction of the maps and reports; Bob Dubasz and Jim Richard for their interest and help as field assistants.

PREVIOUS WORK

Early investigation and mapping of the surficial geology of New Brunswick was initiated by Matthew (1879), and was completed by Chalmers (1884,

1886, 1887, 1888, 1890, 1895, 1901, 1902), who published surficial geology maps for the complete province. Chalmers' concept of local ice caps based on his many observations of striae is still basically valid with minor modification: his maps of the surficial materials are very general and details from inaccessible areas are lacking, probably because aerial photographs were not available to him.

Recent detailed mapping of surficial geology in New Brunswick has been confined to the Saint John River valley (Lee 1955, 1957, 1959a,b, 1962), in areas of urban development (Melvin 1966), and in conjunction with granular resource inventory (*cf.* Brinsmead 1974a,b,c,d, 1977, 1978; Brinsmead and Finamore 1977; Finamore in press; Finamore and Lohse 1977; Hamilton 1973; Kingston 1974a, b; Lohse 1974a,b,c, 1975; Thibault 1978, 1980). Gadd (1973) mapped the Quaternary geology of Fredericton (21 G) map area at a scale of 1:250,000; and Gauthier (personal communication 1980) is mapping the Quaternary geology of Campbellton (21 O), and Bathurst (21 P) and the New Brunswick part of the Edmundston (21 N) map areas at a scale of 1:125,000. This report and map at a scale of 1:250,000 is one of three reports and maps of the Quaternary geology of Amherst (21 H), Moncton (21 I), and Woodstock (21 J) map areas, which will complete the Quaternary mapping of New Brunswick at a scale of 1:125,000 to 1:250,000.

Soils (Aalund and Wicklund 1949, Stobbe 1940) and peatland (Korpijaakko 1976, New Brunswick Department of Natural Resources 1979) maps and investigations deal with regional, but special aspects of the surficial deposits. Numerous studies have been completed dealing with special aspects of the Quaternary geology of New Brunswick (*cf.* Gauthier 1978, 1979, 1980a,b; Gauthier and Thibault 1980; Grant 1970, 1977; Greiner 1974, Kiewiet de Jonge 1951; Mott 1975; Ruitenberg *et al.* 1976, Schroeder and Arsenault 1978; Welstead 1971). Casual observations regarding unconsolidated deposits and glacial history were often made during the course of bedrock investigations (for a bibliography of publications concerning bedrock see Carroll 1977, Poole *et al.* 1970, Ruitenberg *et al.* 1979).

PHYSIOGRAPHY, BEDROCK AND DRAINAGE

The area lies within the New Brunswick Highlands except for the northwestern and northeastern corners, which lie within the Maritime Plain (Bostock 1970). The New Brunswick Highlands that lie east of the Saint John River have been designated as the Caledonian Highlands by Weeks (1957).

The Caledonian Highlands can be separated into two main parts (Fig. 1): the central plateau, which is underlain primarily by Precambrian or lower Paleozoic sedimentary and volcanic rocks and their metamorphosed equivalents, and by a variety of igneous rocks, some of which are Devonian in age (Potter *et al.* 1979); and the boundary area, which is underlain primarily by Mississippian and Pennsylvanian sedimentary rocks with local areas of older sedimentary and metamorphic rocks (Potter *et al.* 1979). The latter are folded and faulted into broad synclines and anticlines.

The central plateau of the Caledonian Highlands flanks the Fundy coast from Alma to St. Martins: its northwestern edge is defined by a line running from Saint John to Prosser Brook: its northern end by the upper course of Weldon Creek. A gently rolling plateau-like surface with local clusters of rounded hills characterizes the topography; its surface declines gently at its southwestern end. The plateau is moderately well drained except locally on flat areas and in depressions. The drainage pattern is predominately radial with streams showing mainly dendritic patterns and being deeply incised at the edge of the plateau surface. At the southwestern end of the plateau the drainage has been glacially deranged and shows an irregular pattern: streams here are less incised than those to the northeast.

Much of the boundary area of the Caledonian Highlands is characterized by ridge and valley topography typical of the Appalachians. The ridge and valley topography, which closely relates to the main structural trends, has northeast-southwest orientation. It is most dramatically illustrated along the

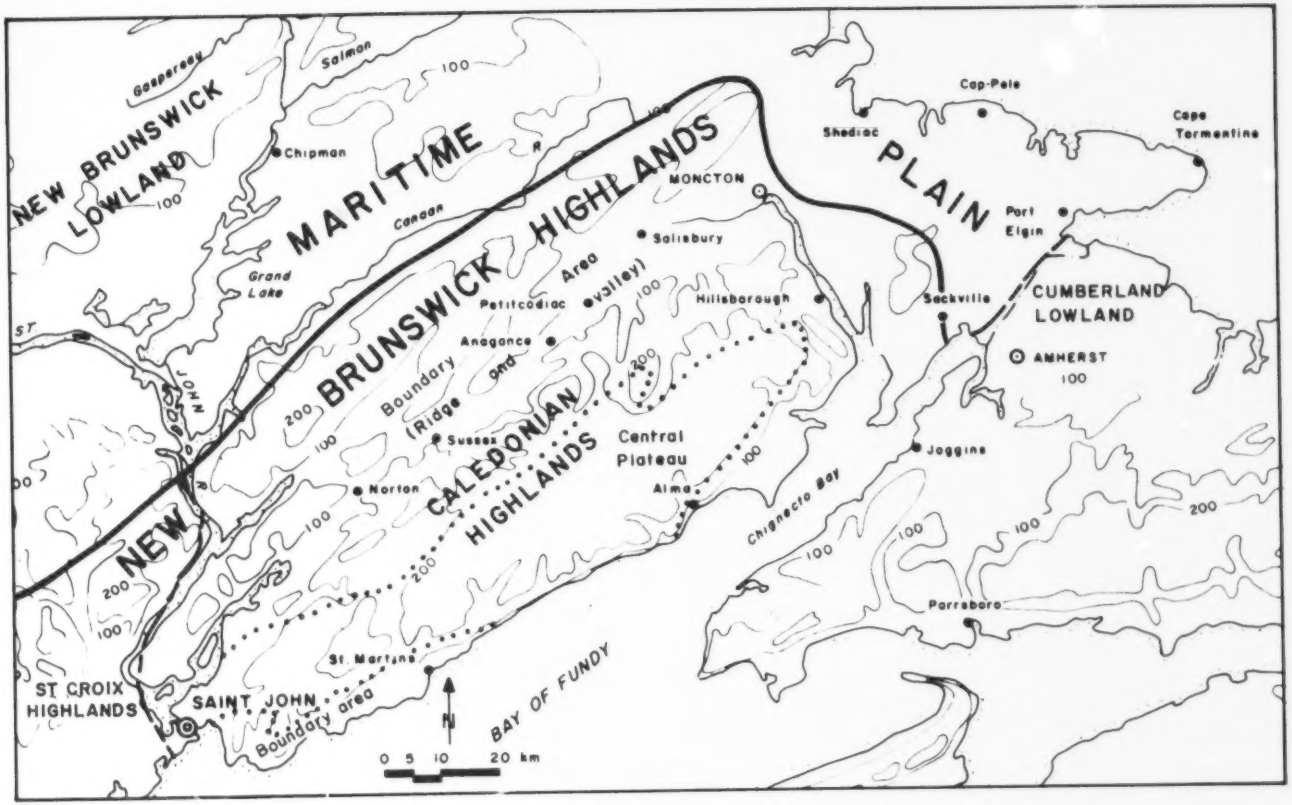

Figure 1. Physiography of southeastern New Brunswick.

northwest edge of the central plateau, and southeast of Riverside-Albert where parallel ridges and valleys have a relief of up to 250 m (800 ft). Drainage is generally good in this area of high relief except in some broad valley bottoms clogged by glacial drift or filled with alluvium and intertidal sediments. The drainage has a trellis pattern characteristic of ridge and valley topography. The major streams such as the Petitcodiac, Anagance, Kennebecasis, Millstream, and Hammond Rivers and Smith and Belleisle Creeks parallel the structural trend over most of their course. Tributaries are both obsequent in that they are at right angles to the main structural trend and consequent in that they parallel structural trends. In areas where the ridge and valley topography is not well developed such as between Petitcodiac and Sackville, glacially deranged drainage patterns reflect regional slopes and range from parallel to dendritic to irregular.

The Maritime Plain (or Carboniferous-Triassic Lowlands) flanking the north edge of the Caledonian Highlands is called the New Brunswick Lowland (Weeks 1957, Williams *et al.* 1972). The New Brunswick Lowland is underlain by Pennsylvanian sedimentary rocks except for isolated outcrop areas of other ages (Potter *et al.* 1979). It is relatively flat to gently sloping with maximum local relief of about 60 m (200 ft) because of the relatively horizontal to gently dipping strata underlying it. Drainage is only moderately good to fair because of its flatness and because Quaternary events have affected the area (glaciation and marine submergence). The drainage pattern is irregular due to these same events.

SURFICIAL GEOLOGY

The effects of weathering, mass wastage, glaciation, marine and lacustrine submergence, isostatic rebound, paludification, alluviation, and stream and coastal erosion, in the past and present, have led to the development of numerous landforms composed of a variety of materials. The landforms and materials resulting from these activities are shown on the accompanying map (Map 81-003).

Description of Map-Units

Units on the accompanying map are described by a series of letters. In this system the letters designate the compositional-genetic category, texture and morphology or thickness of a map-unit. The letter designations can quickly be memorized as they are consistent on all maps regardless of the number of map-units.

Compositional-genetic categories indicate the genesis of a map-unit, and each genesis connotes a possible variety and arrangement of landforms and variety of textural variations within the landforms, both laterally and with depth. For example, 'L' designates that a unit contains materials that are lacustrine in origin, that are probably sand or silt, less probably clay or gravel, that are usually horizontally stratified and coarsen towards their margins and that have a planar surface morphology or a morphology reflecting the shape of the surface on which they were deposited. A textural prefix and morphologic suffix indicate the texture and morphology where they are known with a degree of certainty. For example, 'sLp' indicates a lacustrine unit containing mainly sandy sediment and having a planar surface morphology.

Compositional-genetic notations and categories employed are: A - Alluvial, C - Colluvial, G - Glaciofluvial, L - Lacustrine, M - Morainal, O - Organic, R - Bedrock and W - Marine. Textural prefices and their definition are: a - in most cases sand and gravel, but where used with M, denotes stoney (pebble-sized and larger clast content greater than 35 percent) till; b - bouldery (boulder-sized clast content greater than 25 percent); f - fine-grained (clay, silt, fine sand, minor peat and organic sediment); g - gravel; s - in most cases sand, but where used with M, denotes sandy till; x - complex, usually a mixture of silt, sand, gravel and till. Morphology and thickness

suffixes and their definition are: b - blanket, generally 0.5 to 1.5 m thick (morphology reflects relief of underlying bedrock surface); h - hummocky (numerous moderately steep and steep slopes present; m - rolling (generally gentle to moderate slopes); n - nondescript sloping (morphology reflects relief of underlying bedrock surface) and generally greater than 1.5 m thick; p - plain (includes floodplain); r - ridge (includes beach, bar, and spit); t - terrace; x - complex combination of landform morphologies; v - veneer, generally discontinuous and less than 0.5 m thick (morphology reflects relief of underlying bedrock surface).

Bedrock (R)

This unit includes areas where rock is frequently present at or near the surface. Scattered areas of thin unconsolidated deposits, mainly weathered bedrock and glacial drift on upland surfaces and colluvium on slopes, may be present. In swales and depressions unmapped pockets of thick unconsolidated sediments may be present.

Areas mapped as bedrock are mainly restricted to two areas: (1) ridges and uplands in the boundary area of the Caledonian Highlands, and (2) steep slopes developed by wave and stream erosion in the Caledonian central plateau. In the boundary area west of the central plateau, many upland ridges have only thin drift veneers, but only those areas where bedrock is most frequently exposed have been mapped as bedrock. Isolated outcrops are numerous throughout the boundary area except in valley bottoms filled with thick drift, alluvium or intertidal muds. On uplands the frequency of rock outcrops increases from areas that have been mapped as morainal blanket to areas that have been mapped as morainal veneer.

The naturally exposed bedrock surfaces generally show the effects of weathering although a few surfaces still show the effects of glacial erosion and scour. Weathering may take the form of fracturing and disintegration, or a chemical and physical alteration of the surface. The depth of effective weathering and bedrock disintegration varies from negligible to shallow in areas that have been glaciated during the late Wisconsin to variable, but frequently deep, in areas that may not have been glaciated since the early or middle Wisconsin. In the Caledonian central plateau, local areas of deep rock disintegration are present. Precambrian volcanics near Londonderry (Ruitenberg *et al.* 1979) are broken, disintegrated, and are being excavated for aggregate to depths exceeding 8 m. Weathered and disintegrated sandstone on Pleasant Mountain is thicker relative to similar rocks in adjacent areas. Whether this deep weathering is a function of subtle differences in rock composition or of the glacial and erosional history is not clear. Chalmers (1890, 1895) identified many areas of weathered rock in the area and believed that they were of preglacial origin and were left unscathed by glaciers. This explanation is possible where the bedrock was weathered to great depth such as north of Millville (Rampton and Paradis 1981a) or near Londonderry, but where the bedrock is weathered to a lesser depth at other localities the weathering may well be postglacial.

Bedrock areas commonly show the effects of glacial erosion. In areas where glaciers were moving nearly parallel to the bedrock structure, such as near Mispec and New Horton, the rock is strongly fluted due to the varying erodibility of the different lithologies. In areas where the glaciers were moving at an acute angle to the bedrock structure, such as southwest of Norton, the surface is planated and crudely bossed. On a smaller scale, the bedrock surface is often grooved and striated. Two or more sets of striae having different directions are superimposed on many outcrop. Pebbly conglomerates can show good small scale crag-and-tail.

Undifferentiated Deposits and Landforms

Colluvium (C)

Areas mapped as colluvium are mainly restricted to areas where stream erosion has produced the required slopes and where rock weathering has allowed materials to move down-slope through mass movement (generally slow creep), or where rock weathering has produced a relatively continuous regolith that mantles the underlying bedrock surface and may be subject to little downslope creep. Colluvium is present on other slopes, but is generally indistinguishable from glacial drift (from which it may be derived) or is in patches too small to be mapped.

Material mapped as colluvium consists primarily of weathered and disintegrated local bedrock, but may contain lenses and layers of glacial drift derived from adjacent uplands. Bedrock outcrops may be scattered throughout the map-unit, especially on steep slopes where the colluvium is only a thin blanket or veneer. At the base of many slopes the colluvium thickens and interfingers with alluvium.

The area mapped as colluvial veneer at the north edge of the central plateau between Penobsquis and Baltimore is unique in that it appears to consist of a thin layer of sandstone fragments and very sandy regolith. Rare exotic clasts in this regolith raise the possibility that it may be in part morainal, and that till composed primarily of local sandstone clasts has weathered to a very sandy material.

Most colluvium is probably postglacial in age, but some may have formed during earlier parts of the Quaternary. Generally colluvium on steep slopes can be expected to be younger than colluvium on gentle slopes due to the more rapid downslope movement and erosion on steep slopes.

Glacial Deposits and Landforms

Morainal Deposits (M)

Morainal deposits usually consist of ablation or lodgement till, but often contain glaciofluvial deposits, and even glaciolacustrine deposits. The term "morainal deposits" rather than the term "till" is favoured as a description of these deposits because the complex inter-relationship of ablation till, lodgement till, glaciofluvial deposits and glaciolacustrine deposits are often not subdividable, especially at the scale of mapping undertaken herein.

Morainal deposit map units may contain much weathered bedrock and colluvium. In many localities where the thickness of unconsolidated sediment over bedrock is relatively thin it is difficult to separate bedrock that has been weathered, shattered, frost-heaved and mixed with whatever surficial materials were present, from morainal deposits, which contain clasts that have been weathered and shattered. This is particularly true where morainal deposits contain a superabundance of local lithologies, which break to angular clasts, because the weathered till will have a rubbly appearance. The texture of surface tills may become significantly finer in areas where the clasts are mainly shale (Leavitt and Perkins 1935).

Identification of ablation and lodgement till is difficult, especially in areas where the till is relatively thin and may be affected by soil processes throughout its total thickness. Generally in areas mapped as being covered by lodgement till (Mp, Mn, Mb and Mv), the till in the upper 0.1 to 0.5 m is relatively loose and may be ablation till. However even where underlying till appears compact, this compactness may be due to postglacial subaerial processes acting on ablation till. Gadd (1973) has suggested that much of the loose, poorly-sorted gravelly deposits on morainal surfaces in southwestern New Brunswick may be due to soil-forming processes and slope erosion by running water. Although slope wash, rill wash, and sheetwash (cf. Lee 1962) were active both during and after deglaciation most loose surface till, above limits of submergence by lacustrine or marine waters, is probably a thin mantle of ablation till or lodgement till affected by soil-forming processes, rather than material of a separate origin.

Lodgement Till: Areas showing little evidence of hummocky, rolling or ridged morainic topography, and containing few ice-contact deposits have been mapped as lodgement till (Mp, Mn, Mb, Mv). The indicated thickness of the lodgement till is based on the observed average thickness of till, frequency of bedrock outcrops along ditches and estimates made from geomorphologic properties observed on aerial photographs. Areas mapped as morainal veneer (Mv), which include most of the area west of the central plateau and north of Shepody Bay, generally have a discontinuous till mantle of less than 0.5 m thickness with frequent bedrock outcrops and patches of till 0.5 to 1.5 m thick, and a few pockets of other unconsolidated deposits and till greater than 1.5 m thick. Areas mapped as morainal blanket (Mb) generally are covered by 0.5 to 1.5 m of till, but include some areas where till is less than 0.5 m thick or greater than 1.5 m thick. Occasional bedrock outcroppings and a few patches of other unconsolidated materials may also be present. Areas mapped as morainal plains, either in the form of a plain (Mp) or reflecting underlying bedrock topography (Mn), generally are covered by more than 1.5 m of till although in some areas the till is between 0.5 and 1.5 m thick, and in a few areas till may be less than 0.5 m thick and scattered pockets of other unconsolidated deposits may be present. Thick lodgement till is prevalent in areas near ice marginal positions, such as between Petitcodiac and Stoney Creek and north of Midgic.

The texture and colour of lodgement till is usually closely related to the underlying bedrock, especially where the till is thin and has been mainly derived from the underlying bedrock. Shales tend to produce clayey silty tills, whereas interbedded shales, sandstones, conglomerates and limestones tend to produce a till with more normal clay/silt/sand grain size ratios. Till underlain by sandstone tends to be very sandy in texture, whereas areas underlain by coarse conglomerate or igneous or metamorphic rocks tend to be stoney, bouldery or locally sandy. Tills in near surface exposures are usually red, reddish-brown, brown or yellowish-brown; at depth where the tills are not oxidized they may be grey. Chalmers (1890) reported a bluish-grey till at depth near Alma. The brilliant reddish hues (2.5YR3/6, 10R3/6) are most prevalent where Mississippian, Pennsylvanian or Triassic red shales or sandstones are present, such as south of Sussex and along the northern and eastern edge of the Caledonian Highlands. In a few localities reddish till derived from Mississippian bedrock to the north has been deposited on the northern edge of the Caledonian igneous-metamorphic complex. Along the eastern edge of the Caledonian Highlands the change in colour from red to reddish-brown (2.5YR4/4) or brown (7.5YR4/4) and a textural change from silty to stoney till marks the boundary between till deposited by ice moving southeast across the Caledonian central plateau and till deposited by ice moving southwest toward Chignecto Bay. As a result of ice moving southward out of Northumberland Strait till in the Sackville district is also very red. Small areas of reddish till indicate local outcropping of red shale. Brown hues are common in areas where till is derived from underlying greenish or greyish sandstones, volcanics, metavolcanics, metasediments and igneous rocks.

The Caledonian central plateau has been mapped as a blanket of stoney lodgement till (aMb). The till varies from sandy to rubbly to stoney to bouldery and areas of ablation till may also be present, but the different phases of till could not be subdivided at the scale of this investigation. In addition it is difficult to distinguish weathered and shattered bedrock from weathered tills in some areas. The till thickness varies from negligible in areas of weathered bedrock to greater than 1.5 m in other areas. It also does not allow mapping at the scale of this investigation. However, most frequently the central plateau is covered by till with an average thickness of 0.5 to 1.5 m.

Ablation Moraine: This material is confined mainly to valleys marking significant ice marginal positions west of the central plateau; to the area near McFadden Lake on the central plateau; and to an area near Sackville. The topography of ablation moraines (Mm, Mh) west of the central plateau is rolling with only a few irregular hills and ridges (cf. Brinsmead and Finamore 1977); the topography of this map unit near McFadden Lakes is one of low parallel ridges; and the topography near Sackville varies from rolling to sharply hummocky with many knolls and hills having moderately steep slopes.

Texturally, ablation till has been well described by Brinsmead (1977, p. 13) as consisting "of poorly sorted to largely unsorted material, usually containing, in significant percentages, component particles from the entire range of clast sizes from clay through boulders. It is moderately to poorly compacted, the compaction factor normally being less than that of lodgement till, but may include pockets of better sorted ice-contact granular material". Ablation moraine appears to contain a greater abundance of cobbles and boulders than morainal deposits composed of ablation till: for example, morainal ridges near McManus Lake mapped by Brinsmead and Finamore (1977) are covered by large conglomeratic boulders, and the morainal/outwash complex at the south end of Loch Lomond has a high proportion of boulders. Ablation moraines northeast of Sussex not only contain frequent occurrences of glaciofluvial deposits, but also clay and silt of glaciomarine or glaciolacustrine origin. Glaciofluvial deposits associated with the ablation moraines near Sackville are very sandy as is the ablation till itself in this area. The ablation moraines at McFadden Lake may have in part an ice-thrust origin with bedrock having been incorporated into at least one ridge.

Glaciofluvial Deposits (G)

Glaciofluvial deposits include those sediments deposited by meltwater in contact with glacial ice (ice-contact stratified drift: Gr, Gh, Gm), and proglacially (outwash: Gp, Gn, Gd, Gb, Gv). In many areas subglacial and englacial meltwater channels remained the focus of meltwater flow following the retreat of glacier ice from the locality; thus ice-contact stratified drift and outwash are often found in close association (Gx).

Clasts within most glaciofluvial deposits generally reflect the lithology of nearby bedrock or of rock immediately to the north and west. However, east of Sussex, an inordinate number of clasts in the glaciofluvial deposits originated from the Caledonian central plateau, and suggest that older alluvium deposited by streams flowing west from the central plateau has been incorporated. Indeed, some deposits mapped as glaciofluvial along the west and north edge of the central plateau may have been deposited by streams flowing out of the central plateau, and aggrading to levels controlled by ice dams in valleys.

A detailed description of glaciofluvial deposits is given for individual deposits over much of the area by Brinsmead (1974a,b, 1977), Brinsmead and Finamore (1977), Finamore and Lohse (1977), Lohse (1974a) and Thibault (1978).

Ice-Contact Deposits: Eskers and esker-like ridges (Gr), kames and kame-and-kettle complexes (Gh, Gm) are the main landforms in which ice-contact stratified drift is present. Generally these landforms are closely associated with each other or with ablation moraines.

Ice-contact stratified drift "characteristically comprise well- to poorly-stratified sand and gravel that may include varying amounts of silt and clay. Ice-dropped boulders are common in this material.

Stratification of ice-contact deposits is commonly contorted and faulted..... Abrupt changes in grain size and degree of sorting are characteristic of such deposits as well." (Brinsmead 1977, p. 14).

Characteristically, ice-contact deposits contain layers and lenses of till or till-like material. Often the latter mantles the stratified drift and as such may be a capping of ablation till sloughed off or let down by melting ice on subglacially or englacially deposited stratified drift. This till-like material is poorly-sorted, bouldery, but shows signs of at least partial water washing and removal of some fines.

Ice-contact deposits vary in size and many are too small to be mapped at the scale of this investigation. Significant ice-contact deposits are present at the following locations: (1) Walton Lake and the head of the Big Salmon River drainage where a complex of kettled gravelly outwash, and esker ridges are present along a major meltwater channel crossing the central plateau; (2) near Barnesville where a large esker complex leading from the Hammond River through a bedrock gap terminates in a kame-and-kettle complex near Third Lake at the head of the Loch Lomond system; (3) in the area surrounding Haynes

and Taylor Lakes (Finamore and Lohse 1977); (4) in the Little River valley (cf. Melvin 1966) where the relatively bouldery gravels near Ben Lomond grade to sandier stratified drift southwest of Lakewood (these deposits probably mark a former ice marginal position); (5) north of Golden Grove near Bradley and MacFarlane Lakes (cf. Melvin 1966) where a complex of esker ridges, kames and outwash is composed mainly of gravel and sand, but contains some silty sand and stratified sand, silt and clay. The latter were probably deposited in a glaciolacustrine or glaciomarine environment; cf. Finamore and Lohse 1977); (6) east of Sussex to near Five Points where sand and gravel are associated with till in an area of low rolling ablation moraine (cf. Brinsmead 1977); and (7) between Dubee Settlement and Belleisle Creek (cf. Brinsmead and Finamore 1977) where numerous esker-like ridges and other ice-contact forms appear to have formed by meltwater flowing southeast into the Belleisle/Millstream valley and southwest parallel to the valley through stagnant ice occupying the valley.

Outwash: Valley trains (Gp) and glaciomarine deltas (Gd) constitute most of the outwash in the area. Where the surface of the outwash is not particularly flat, but reflects the topography of the underlying bedrock it has been mapped as Gn; where it mantles bedrock and averages between 0.5 and 1.5 m in thickness it has been mapped as blanket (Gb); and where it mantles bedrock and is generally less than 0.5 m thick it has been mapped as veneer (Gv). The latter three map units may contain some ice-contact stratified drift, but because of the lack of any landform morphology that would systematically identify it as such no separation has been attempted.

Brinsmead (1977, p. 15) has described outwash in general as being "typically well sorted and fairly well stratified gravel and sandy gravel. The content of fine particles is usually low in outwash deposits, as the meltwater from which the coarse material is deposited tends to carry the fines farther downstream.", but qualifies deltaic deposits as follows: "The base of a typical section comprises fine to silty fine sand with silt interbeds, grading upwards through medium sands and pebbly sand into medium-pebble to small-cobble gravel near the surface..... Stratification is much more regular and laterally continuous than that of outwash, which is more likely to exhibit cut-and-fill structures and truncated bedding." (Brinsmead 1977, p. 16).

Significant outwash deposits are present along the Fundy coast from Black River to the Petitcodiac estuary (cf. Brinsmead 1974b, Finamore and Lohse 1977, Thibault 1978), in the upper drainage system of the Big Salmon River along the major meltwater channel crossing the central plateau; south and west of ice contact deposits and ablation moraine at Loch Lomond and in the Little River valley; along the Kennebecasis River above Penobsquis, and along Smith Creek above Newtown (cf. Brinsmead 1977); and within the glaciofluvial complex at Elgin (cf. Brinsmead 1974a).

Outwash deposits along the Fundy Coast appear to be mainly deltaic in origin, and have been mapped as such (Gd) at Black River, and at Lower Cape. In cases where they may be in part deltaic and in part outwash plain or valley train such as near St. Martins, Waterside, and Stoney Creek they have been mapped as outwash (Gp).

Outwash deposits along Smith Creek and the Kennebecasis River are similar in that they are apparently of deltaic origin, being deposited either in an estuarine environment (Brinsmead 1977) or in lakes dammed by ice in the lower parts of the valley. The deposits along Smith Creek are further complicated by the occasional presence of a till layer within them; and complicated along the Kennebecasis River by the fact that the river may simply have been grading to an ice front under which it drained subglacially.

Late Glacial and Postglacial Deposits and Landforms

Submerged and Wave-Washed Surfaces

Characteristically, areas indicated as having definitely or probably been submerged or wave-washed have a discontinuous veneer of sand over the underlying material. The sand veneer may contain pebbly and silty layers.

Pockets and lenses of silt and clay are present in some swales and environs protected from former intense wave-wash; and patches of sandy to pebbly gravel on hillsides and hilltops exposed to former intense wave-wash. Locally small areas of late glacial lacustrine and marine sediment thicker than 0.5 m are present.

Two phenomena during the deglaciation of New Brunswick have led to the wave-washing and submergence of large areas: postglacial submergence by the sea due to isostatic depression of the land by glaciers and the ice damming of valleys by retreating glaciers. Whether the resulting deposits have a lacustrine, glaciolacustrine or marine origin is often difficult to determine as they are almost universally devoid of fossils, which would allow a determination of environment of deposition. The absence of marine fossils in eastern New Brunswick is probably due to a variety of phenomena: (a) the proximity of waning glaciers supplying large quantities of relatively cold fresh turbid water, (b) the relative shallowness of the waters covering part of the submerged area, (c) a land-sea configuration consisting of long estuaries with relatively fresh water along much of their length, and (d) the probable destruction of calcareous fossils because of the general absence of lime to act as a buffering agent, and the permeability of the porous sands and sandstones over most of the Maritime Lowland (Chalmers 1895, 1902).

The extent of area shown as being submerged and wave-washed was determined by a number of criteria: (1) deposits containing marine shells or clay of probable marine origin at similar or higher elevations; (2) a relatively continuous veneer or blanket of silty sand or sand with pebbly layers over other materials (especially where the underlying material is a compact silty or clayey till, which would not easily produce a sandy soil horizon upon weathering); and (3) the absence of distinctive sharply defined glacial landforms on aerial photographs. In the last case hummocky or ribbed moraines were often found during field checking, but steep slopes were generally absent indicating that either the feature developed with gentle slopes due to subaqueous formation or had been modified by submergence and wave-washing.

Areas where evidence indicated a high, but not definite, possibility of submergence have been shown on Map 81-003 as having probably been submerged and wave-washed. The ultimate limit of submergence was often difficult to determine for several reasons: (1) sand or silty sand does not universally overlie the non-marine deposits, even in areas of known submergence, (2) a sand or silty sand mantle, if present, could have had an alternate origin, e.g. glaciofluvial or alluvial, (3) disturbance by agricultural and logging activities has destroyed the evidence, and (4) many of the areas where this limit probably occurs are inaccessible.

Most areas indicated as having been submerged and wave-washed were probably affected by marine water, or at least water below the elevation of sea level. However it is possible that parts of the Belleisle Creek/Millstream River, Kennebecasis River, Hammond River, Anagance River/Petitcodiac River valleys were affected by proglacial lakes. Areas shown on Map 81-003 as having only probably been submerged and wave-washed in the upper courses of the Hammond River, Pollett River and other streams draining the northwest edge of the Caledonian Highlands could only have been submerged by the water of ice dammed lakes. The extent of submergence and wave-washing is derived from the position of assumed lake outlets.

Areas of high elevation north of Berryton and Dawson Settlement that appear to have been submerged are well above the regional maximum marine limit of submergence and are located such that lakes could not be dammed by ice blocking downstream ends of valleys. A lake may have formed in this area after it had been deglaciated, while ice remained to the east in Chignecto Bay, to the north near Moncton and to the west near Petitcodiac.

Marine and Lacustrine Deposits (M,L)

Late glacial or early postglacial deposits that appear to have been deposited in marine or lacustrine environments have generally been mapped as marine, even though fossiliferous material that would identify them as such are lacking. This is because most such deposits lie below apparent marine

limits, and the fine-grained facies generally lacks distinctive varves characteristic of a glacial lake. Nevertheless, some deposits may well have been deposited in drainages blocked by glaciers and fed only by streams and meltwater. Most other investigators present a similar analysis of the situation (cf. Brinsmead 1977, Brinsmead and Finamore 1977).

The few marine deposits that have been mapped are near St. Martins, Red Head, Model Farm and Smithtown (cf. Finamore and Lohse 1977) near Norton (cf. Brinsmead and Finamore 1977) and Canaan Rapids. Typically, where the deposits are thin, averaging 0.5 to 1.5 m in thickness (Wb), they usually contain a high proportion of gravel and sand, as near St. Martins and Canaan Rapids. Where the deposits are over 1.5 m thick (Wp), they usually consist mainly of clay, silt and fine sand, in places with interbeds of coarse sand or pebbly gravel. Near Model Farm 7.5 m of marine or estuarine (or lacustrine) silty sands overlie gravel and sand of probable glaciofluvial origin, and near Smithtown 2 m of interbedded clay, silt and sand are overlain by a metre of marine sand and gravel (Finamore and Lohse 1977). At Norton, a complex of supposed shallow and deep water deposits exist that consist mainly of interbedded clayey silt and silty sand (Brinsmead and Finamore 1977).

Other deposits of probable marine or lacustrine origin have been identified in the ridge and valley part of the Caledonian Highlands, but they are only present in the subsurface or are too small to map and have been included with glaciofluvial or morainal deposits. Deposits mapped as glaciofluvial deltas (Gd) or outwash (Gp) were in part deposited in a marine environment. For example, deltas and outwash mapped along the Fundy coast, often contain significant layers of marine silt and sand within them. Near Carpenter Pond southeast of Rothesay a complex glaciofluvial sequence contains fine-grained rhythmites of glaciolacustrine or marine origin (Finamore and Lohse 1977, Melvin 1966). Even north of Sussex along the Kennebecasis River and Smith Creek, mapped outwash contain sediment that was probably deposited in an estuarine or lacustrine environment.

Exposures of clay and silt under till southwest of Penobsquis, in stream-eroded terraces at Sussex and in the Petitcodiac River valley near Petitcodiac (Brinsmead 1974a) indicate that significant thicknesses of marine silt and clay probably underlie glacial drift and alluvium in many valleys of the ridge and valley area at low elevations.

Postglacial Deposits and Landforms

Organic Deposits (O)

Organic deposits are mainly restricted to peat bogs, which are located in depressions and swales southeast of Loch Lomond and north of Snider Mountain. Small unmapped bogs are present in hollows and depressions throughout the central plateau and on the flat parts of the boundary area toward Grand Lake and east of Petitcodiac. Shallow areas of peat may also be present within postglacial marine sediments north of the Cumberland Basin (cf. Aalund and Wicklund 1949, New Brunswick Department of Natural Resources 1977).

Most large bogs probably contain peat exceeding one metre in thickness, but small bogs may be underlain by thinner peat. The upper part of most peat bogs consists of *Sphagnum* peat with a variety of peat-types and mucks interlayered at depth (cf. Korpijaakko 1976). Silt, clay and sand may be interbedded with the peat and muck where lakes have been infilled. Most organic deposits in New Brunswick appear to have formed during the last 10,000 years (cf. Korpijaakko 1976).

Alluvial Deposits (A)

Only relatively broad floodplains and terraces could be shown on Map 81-003. Narrow floodplains and terraces however line most stream courses. An attempt has been made to separate those valleys where terraces (At) predominate from those valleys where floodplains (Ap) predominate, even though most streams have both landforms bordering them. Low flats adjacent to streams

and floodplains that show no signs of constant flooding have been classified as terraces, even though they may be occasionally flooded (e.g. by the one in a hundred year flood).

Some areas mapped as alluvial terrace (At) near Sussex have been shown as outwash by brinsmead and Finamore (1977). Although these terraces most certainly contain eroded glaciofluvial (and marine materials), their geographic location and slope seem to indicate an alluvial origin.

Gravels, frequently cobbly (and bouldery), are the predominant sediment in the alluvium along the upper courses of most streams. Brinsmead (1977) has observed that terrace gravels are often pebbly to cobbly and cleaner than glaciofluvial deposits from which they are reworked. Downstream the alluvium generally becomes sandier and siltier, a function of finer-grained source material and lower stream gradients. Most alluvial gravel and sand have a thin cover of silt or silty sand.

The fine-grained material along the lower courses of the Kennebecasis River and Belleisle Creek is considered alluvial because of the alluvial landforms (meandering channel, levees, etc.) that characterize them. Undoubtedly, these areas are affected by tides, and some sediment may be deposited from brackish water. The sediment underlying these areas is probably interbedded sand, silt and clay with variable organic contents. Streams draining to the Fundy coast have valleys whose lower courses are more clearly filled with marine (intertidal) material.

Marine Deposits (W)

Large expanses of intertidal deposits constitute the Tantramar, Hopewell, Petitcodiac and Memramcook marshes at the head of Chignecto Bay. Their poorly drained surfaces are relatively flat, and only interrupted by lakes and tidal channels (Fig. 2). Presently, most salt marshes are protected from flooding by dykes and dams. These deposits are mainly silty clays, although thin peaty beds can be present throughout their total thickness. The organic content of the clay is often variable, probably a function of the rate of accumulation of the intertidal deposits and the rate of coastal submergence. The intertidal sediments are frequently more than 10 m thick, and may exceed 30 m in some low areas (Welstead 1971). The thick sequence of intertidal muds is due to coastal submergence and the erosional and depositional actions of the Fundy tides (Chalmers 1895).

Beaches are narrow along most of the Fundy coast because of the steep rocky shorelines and high tidal range. However, bay-mouth bars and spits are often present adjacent to the mouths of streams with steep gradients draining the southwest edge of the Caledonian Highlands or where glaciofluvial deposits are being eroded. The bay-mouth bars generally consist of pebbly or sandy gravel, but may be cobbly and littered with drift wood.

INTERPRETATION OF QUATERNARY HISTORY

Introduction

Most landforms and unconsolidated sediments in this area were formed or deposited during (a) an interval of erosion during which streams deeply incised their courses and developed mature dendritic drainage patterns around the edges of the Caledonian Highlands, (b) a glaciation of the Caledonian Highlands before the interval of erosion, (c) glaciation of the area adjoining the Caledonian Highlands after the interval of erosion, (d) late glacial and early postglacial marine submergence of low lying areas, (e) stream erosion and alluviation following deglaciation and postglacial land emergence, and (f) relatively recent postglacial submergence.

Features and materials, which may in part be Quaternary in age, but which appear to predate the earliest recorded glaciation, are: major valleys such as the Kennebecasis River valley and major scarps such as that forming the

northwest edge of the Bay of Fundy; a low bedrock bench observed west of Rockport; and weathered disintegrated bedrock underlying drift. The low bedrock bench is believed to have been developed by marine erosion. The local pockets of weathered, disintegrated bedrock may have formed, totally or in part, before or after deposition of the overlying drift. Chalmers (1890, 1895) suggests that most rotten rock occurs in areas that have not been subsequently glaciated. The location of some of these areas within regions showing evidence of intense glaciation deems the hypothesis to be at least in part suspect. Casual inspection suggests that disintegrated bedrock is generally thicker in areas that have not been glaciated since the interval of erosion than in areas glaciated more recently. This may indicate that the thickness of disintegrated bedrock is in part related to the length of time it has experienced subaerial weathering.

Early Sea Levels

A marine bench that extends up to 4 m above its modern counterpart has been recorded at numerous localities throughout Atlantic Canada and adjacent areas (Grant 1980a,b). At Rockport and in a cove southwest of Rockport a rock bench is present under unconsolidated materials whose surface stands between 2.5 and 4 m above its modern counterpart at these sites (Fig. 3). In the small cove southwest of Rockport, the surface of the rock bench is intensely striated, but appears to show pothole-like features that characterize wave planated rock benches. A few poorly exposed rock benches were noted at equivalent elevations along the coast southwest of Salmon River, but generally they were not exposed over sufficiently extensive areas for positive identification.

The age of the rock bench could not be closely bracketed in this map area, although interlayered sand and red sandy till overlie the bench at the locality southwest of Rockport. However, even the most recent glacial pulse that transgressed this area (the Chignecto phase) appears to predate 18,000 B.P. and consequently the rock bench is older than this. Grant (1980a,b) has established an age of greater than 38,000 B.P. for planation of this rock bench in other areas of Atlantic Canada through stratigraphic relationships and speculates that it is Sangamon in age.

Early Glaciation

Although most unconsolidated deposits in the area appear to have resulted from Wisconsin or post-Wisconsin processes, earlier glacial intervals probably occurred. Erratics of Laurentide (?) and New Brunswick provenance that are present in Nova Scotia and on the adjacent continental shelf (Grant 1977, 1980b) may have been transported by glaciers of pre-Wisconsin age, having sources west and northwest of New Brunswick.

Wisconsin Glaciations

General Chronology

Glaciations that postdate the construction of the marine rock bench that stands 2 to 4 m above its modern counterpart and is thought to be of Sangamon age, are believed to be Wisconsin in age. In New Brunswick most glaciations and glacial phases are assigned a late or middle Wisconsin age. However, circumstantial evidence exists that the glaciation affecting most of the Caledonian Highlands is early or middle Wisconsin.

Nomenclature

Glacier flow patterns have been established from striae, flutings, and drumlins shown on Map 81-003, and maps from adjacent areas (Gadd 1973, Gauthier 1978, 1979, Rampton and Paradis 1981a,b) and their relative chronologies.

Figure 2. Intertidal sediments exposed along Tantramar River near Sackville.

Figure 3. Wave-planated rock bench southwest of Rockport. Bench stands about 4 m above its modern counterpart. It is marked by superimposed striae and overlain by interlayered till and glaciofluvial sediments. The bank is capped by marine sand and gravel.

Generally, each flow pattern has a distinctive aerial and relative chronologic character that suggests one or more centres of ice accumulation and dispersion. The aerial distribution of flow patterns also often suggest ice marginal positions. Occasionally these ice marginal positions are marked by hummocky, rolling or ribbed moraines.

In further discussion, we have given names to the flow patterns, to morainic belts where the flow patterns terminate, and to ice centres from which the flow patterns appear to originate. The flow patterns and ice margins or glacial limits are interpretative phenomena based on striae patterns, moraines and trim lines, whereas the ice centres are even more hypothetical in that they are based on an interpretation of the flow patterns. In most cases named glacial phases are time intervals during which identifiable flow patterns dominated. In a few cases the glacial phases define short time intervals during which morainal belts were being formed or an ice margin existed.

Source of Glaciers in Southeastern New Brunswick

Southeastern New Brunswick has primarily been affected by glacier ice from two sources: (1) ice originating to the northwest of southeastern New Brunswick and, (2) ice locally accumulating on the lowlands and adjacent continental shelf. Striae suggest that the latter source had many centres of accumulation with the strength and relative importance of each constantly changing (Rampton and Paradis 1981b).

The exact origin of the ice originating west of the map area can only be hypothesized at this time. No erratics having a Laurentide provenance were identified east of the Miramichi Highlands; and perhaps Chalmers (1895), who also made the same observation, was correct in his assertion that glaciers originating to the west of, but crossing eastern New Brunswick, originated on the south side of the Appalachians or Notre Dame mountains. It is also possible that a dome of ice centred on the southern end of the Miramichi Highlands could have been the source for eastward moving ice during some intervals.

Divergent striae patterns indicate that ice caps having different centres of accumulation formed on the Maritime Plain of central New Brunswick and on the adjacent continental shelf during Wisconsin time. The ice caps have been collectively assigned to the "Northumberland Glacier". This concept corresponds in part with Chalmers' (1895) concept of the same phenomenon. However, Chalmers visualized a Northumberland glacier that initially had its source in the Miramichi Highlands, and that only produced divergent flow patterns after it thinned to a point where local topography affected glacier flow. Striae present in southeastern New Brunswick (cf. Rampton and Paradis 1981b) and their chronologic sequence indicates separate, shifting ice centres lying on the Maritime Plain and on the adjacent continental shelf. Striae patterns, drift distribution and tors present on isolated peaks in the Miramichi Highlands (Gauthier 1980b) suggest that the highest parts of the Miramichi Highlands were never a major source of ice during the Wisconsin. In summary, independent local ice centres sporadically affected by pulses of ice from a northwestern source appear to best explain most glacial phenomena in eastern New Brunswick.

Caledonia Phase

The Caledonian Highlands can be divided into three geomorphic zones: (1) a flat plateau area characterized by drift-mantled rolling hills with scattered outcrops showing striated surfaces, (2) a peripheral zone showing deep stream dissection and dendritic stream development, and (3) an outermost edge where tributary stream valleys are less incised and appear partially clogged with glacial debris (Fig. 4 and 5). Striae on the plateau characteristically have a 125° to 155° trend toward its southwestern end and a 120° to 150° trend toward its northeastern end. This trend of glacial flow has been termed the Caledonia flow pattern (Fig. 6). In a number of localities to the east of the Caledonian Highlands, including one site just southwest of Rockport, similar trending striae have been noted and assigned to this flow pattern. Striae with an easterly trend along Chignecto Bay and near Port Elgin to the north (Prest *et al.* 1972) have tentatively been assigned to the Caledonia phase, although

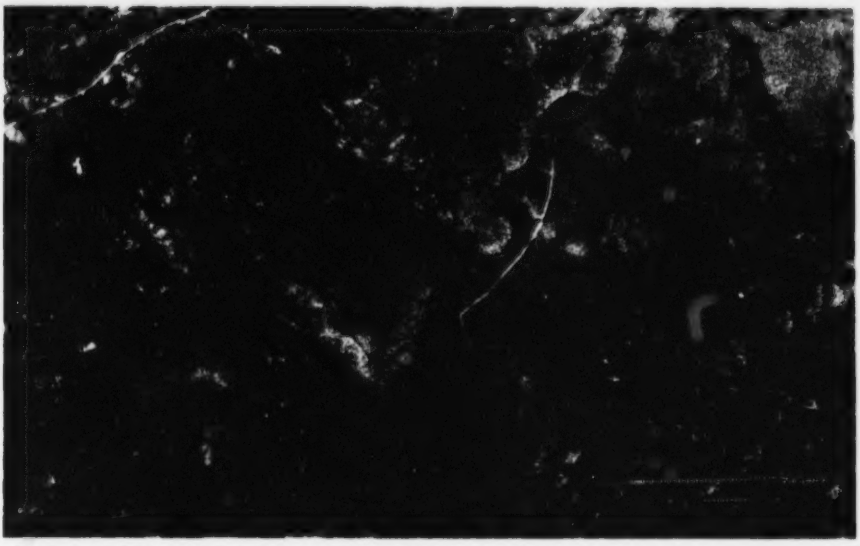

Figure 4. Drainage patterns near South Branch. Dotted line indicates boundary between incised drainage showing mature dendritic pattern to southeast and area where tributary streams are partially filled with post-Caledonia phase glacial drift to northwest (EMR Air photo A21331-72).

Figure 5. Drainage patterns near Berryton. Dotted line indicates boundary between incised drainage showing mature dendritic pattern to south and area where drainage pattern is not well developed due to effects of post-Caledonia glacial phases to north. Note meltwater channel paralleling probable Chignecto glacial limit (EMR Air photo A21331-42).

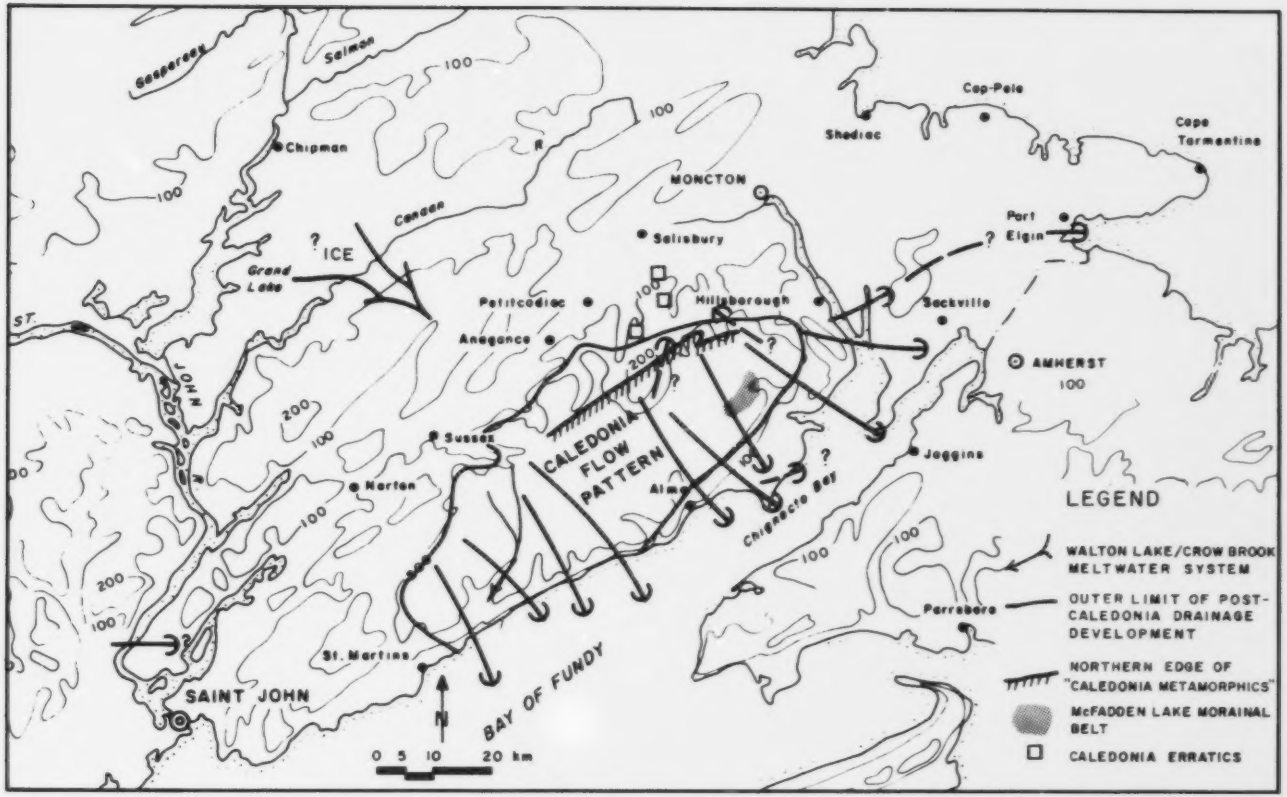

Figure 6. Glacial features of Caledonia phase.

those at Port Elgin could perhaps have been developed during a subsequent phase (Rampton and Paradis 1981b). Rocks with lithologies similar to Precambrian metamorphic and igneous rocks of the Caledonian central plateau are abundant in glacial drift at Rockport and Beaumont.

During most of the Caledonia phase, it appears that ice was flowing from a source area to the northwest across the Caledonian Highlands as attested to by striae generally showing a southeast movement of ice. Fragments of Mississippian conglomerate and sandstone in the drift capping the uplands are also indicative of such glacier flow. However, some indication of northward flow is given by the presence of Precambrian metavolcanic erratics north of the northern limit of bedrock outcrops of this type (Fig. 6). Although, rocks from the Caledonian Highlands could have been transported northward by streams, the erratics indicated in Figure 6 were over 0.3 m diameter and found at interfluve localities. Chalmers (1890) has indicated that ice moved both northward and southward in the vicintiy of Gowland Mountain. However, his direction of ice movement appears to be based on the aspect of outcrops showing glaciated and unglaciated surfaces (Chalmers 1890, p. 45N). Matthew (1879) indicated no evidence of northward flowing ice in this area, and in this survey indications only of southward ice flow on the uplands were found.

Evidence of northward flow along the north flank of the central plateau, northeastward flow near Beaumont, and eastward flow along Chignecto Bay may have been formed during an early part of the Caledonia phase. The extent of ice movements away from the north and east edges of the uplands appears to have been relatively limited because Caledonia-type erratics are rare in the drift to the north toward Salisbury. Their frequency also decreases rather dramatically northeast of the localities at Rockport and Beaumont where they were so abundant in the glacial drift. Eastward flow north of Aulac is suggested by erratics of probable Caledonian origin (Ruitenberg, written communication 1980) which were collected along the coast of Baie Verte to near Cape Tormentine. Fine-grained two-mica granitic erratics, collected at Rockport and north of Aulac resemble bedrock locally present in southwestern and central New Brunswick and in the Cobequid Mountains of Nova Scotia. The latter possible affinity and some questionable nailhead-type striae near Rockport suggest movement off the Cobequids in a northeast direction during the early Caledonia phase.

During the main part of the Caledonia phase an area of ridged and ribbed moraine was constructed near McFadden Lake. Ridges in this morainal belt have a general southwest-northeast trend suggesting that the glacier was flowing from the northwest during their construction. One exposure shows that the local bedrock has been sheared up into the overlying drift. This is the only locality in southern New Brunswick where this phenomena has been noted and it perhpas indicates that a different thermal regime existed at the base of the Caledonia phase glaciers than at the base of later glaciers. Perhaps the Caledonia phase was characterized by a glacier base at sub-freezing temperatures, whereas the bases of later glaciers were near the pressure-melting point.

During deglaciation, eskers were constructed in the Crow Brook valley east of Londonderry and the Walton Lake valley south of Chambers Settlement. As the glacier retreated further north these valleys were occupied by meltwater streams and outwash was deposited along their bottoms. The Walton Lake valley and a valley just south of Saddleback Mountain appear to be the only valleys that may have functioned as spillways for short-lived proglacial lakes along the northwest edge of the central plateau during the waning stages of the Caledonia phase.

At Salmon River on the Bay of Fundy, a delta is present at about 45 m elevation that shows some large-scale foreset beds exposed on its seaward edge. This delta could possibly have been constructed during the waning stages of the Caledonia phase.

Near Rockport striae of the Caledonia flow pattern are present on the marine rock bench, which has been assigned a Sangamon age, and are superseded by striae of later glacial phases, which are believed to have occurred before 18,000 B.P. (see discussion on Chignecto and Kent phases). The Caledonia phase is therefore probably early or middle Wisconsin in age, even though this age cannot be readily substantiated by weathering profiles on the Caledonian drift or radiocarbon dates above or below it.

The weathered and disintegrated nature of surfaces glaciated during the Caledonia phase relative to those glaciated during subsequent glacial phases; and the stream incision and stream pattern development around the edge of the Caledonian Highlands all suggest relative antiquity of the Caledonia phase. Deeply disintegrated bedrock is occasionally present on the surface of the Caledonian Highlands. An extreme example is present near Londonderry where Precambrian volcanics (Ruitenberg *et al.* 1979) are broken and disintegrated to depths exceeding 8 m. Similar rocks to the southwest in areas glaciated subsequent to the Caledonia phase appear less shattered and broken, but this may be simply a function of rock lithology and jointing. The weathered, disintegrated and colluviated sandstone mantling Pleasant Mountain northwest of Elgin also appears relatively thick in comparison to similar sandstones in areas to the north glaciated by subsequent glacial phases. However, occurrences of deeply disintegrated and broken sandstone have been reported from the lowlands to the north (Chalmers 1895). The periphery of the Caledonian Highlands glaciated only by the Caledonia phase and not by later glacial phases is characterized by deep stream incision and m ure dendritic drainage patterns (Fig. 4, 5 and 7). Although this is in part a function of the high relief around the edges of the uplands, part of it may be due to the amount of time since the Caledonia phase. Adjacent areas with similar relief that have been covered by glaciers after the Caledonia phase do not show the same bold stream incision nor the well-developed dendritic pattern of drainage in the headwaters of streams.

Figure 7. Chignecto glacial limit near Weldon Creek. Note arrows indicating meltwater channels, Gm indicating glaciofluvial deposits, Mm indicating rolling moraine constructed during Chignecto or later glacial phases, and incised drainage showing mature dendritic pattern beyond the Chignecto limit (EMR Air photo A21331-43).

Chignecto Phase

Striae and fluting on land bordering the central and northern reaches of Chignecto Bay, that have a south and south-southwest trend have been termed the Chignecto Bay flow pattern (Fig. 8). South trending striae near Shediac have also been designated as part of this flow pattern, and along with other glacial phenomena north of Shediac define a source of ice in northern Northumberland Strait known as the Escuminac ice centre (Rampton and Paradis 1981b). Chalmers (1895) defined "the Chignecto glacier" to account for the striae of the Chignecto Bay flow pattern, but assumed that for the ice to move in this direction the Isthmus of Chignecto and Northumberland Strait must have been elevated at that time.

The edge of Chignecto Bay flow pattern along the northeastern end of the Caledonian Highlands is marked by a change in the trend of striae and fluting from those directions characteristic of the Caledonia phase to those south and southwest directions characteristic of the Chignecto phase. Of this change in flow pattern, a change in colour and texture of till from the brownish and reddish-brown stoney sandy tills of the Caledonian central plateau to the reddish silty tills deposited by the Chignecto phase glaciers exists. In addition, valleys are partially blocked with glacial debris up to the Chignecto glacial limit and valleys are dissected and eroded beyond and above the Chignecto glacial limit. The limit between terrain characterized by stream incision and mature dendritic drainage patterns, and terrain characterized by glacially-deranged drainage and irregular topography due to irregular glacial deposition can be traced around most of the Caledonian central plateau and is believed to mark the upper limit of Chignecto and subsequent Kent phase glaciers (Fig. 8). Only between Hillsdale and St. Martins at the southwestern end of the Caledonian Highlands is this limit difficult to trace. A slight discordancy does however exist between the general flow pattern west of Wood Lake and the Caledonia flow pattern east of Wood Lake; and Salmon River and its tributaries are deeply entrenched. Thus the limit of the post-Caledonia phase glaciers is probably between Wood Lake and the Salmon drainage system.

The striae and flutings to the north and west of St. Martins have been termed the St. Martins flow pattern. This flow pattern and associated phenomena have been tentatively assigned to a later glacial phase (the Kent phase), but they may well have been active during the Chignecto phase. The St. Martins flow pattern would have had its centre of ice accumulation north of Chipman and perhaps stemmed from the Gaspereau ice cap (Rampton and Paradis 1981a,b).

The limit of the Chignecto phase glaciers falls from about 230 m (750 ft) elevation near Albert Mines to about 200 m (650 ft) elevation near Alma, and slopes seaward to the south of Alma. West of Albert Mines, it slopes downward to near 180 m (600 ft) elevation near Pollett River and 150 m (500 ft) elevation east of Penobsquis. However, it appears to be near 230 m (750 ft) elevation south of Sussex. This apparent discrepancy suggests that ice forming the St. Martins flow pattern was probably not synchronous with ice forming the Chignecto Bay flow pattern, and may have developed during a later glacial phase.

During the interval when the Chignecto phase glaciers were near their maximum extent streams draining to the north and west from the Caledonian central plateau would have been dammed, and lakes would have formed as is suggested by sand-veneered drift surfaces in some of these valleys. North flowing tributaries of the Petitcodiac River such as Little River and Turtle Creek have accumulations of outwash or alluvium that suggest these streams graded to the Chignecto ice margin. Few meltwater channels are present to indicate that water was moving laterally between the ice sheet and the Caledonian Highlands. Probably these lakes were ephemeral and drained subglacially or englacially as many lakes in ice dammed valleys do in mountainous regions today.

During the retreat of the Chignecto phase glaciers, kame deltas and glaciomarine deltas formed along its western and southern boundaries (Lohse 1974a, Thibault 1978). A significant glaciomarine delta at Waterside (Thibault 1978) appears to have been deposited when the Shepody River lowland was filled

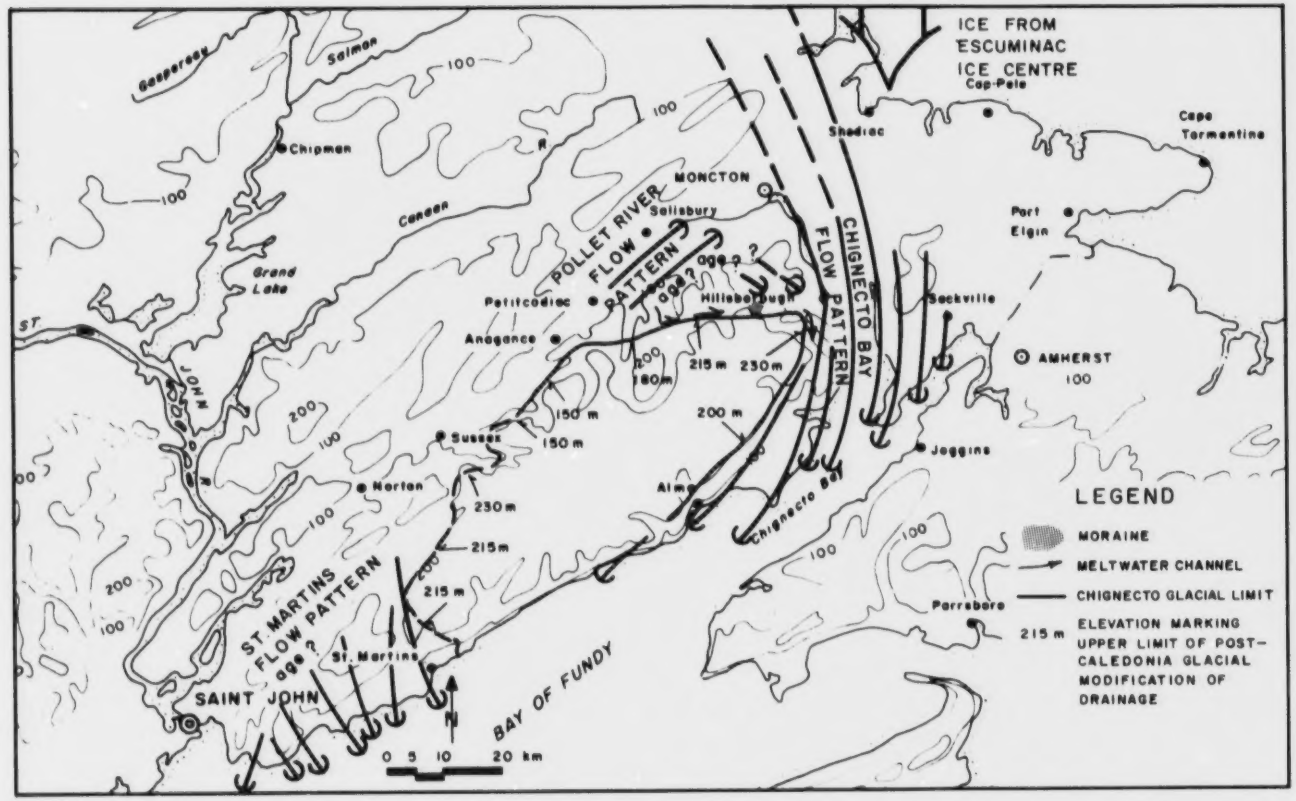

Figure 8. Glacial features of Chignecto phase.

with ice to the north and Rocher Bay was open to the south. Smaller sandy glaciolacustrine deposits, kame deltas and glaciomarine deltas are present along the west edge of Shepody Bay and the Petitcodiac River (Lohse 1974a).

Marine clay near St-Louis-de-Kent contained a whale bone that was dated at 18,570 ± 500 B.P. (S-1969). The marine clay appears to postdate a glacial phase younger than the Chignecto phase (Rampton and Paradis, 1981b). This suggests that the Chignecto phase occurred prior to 18,000 B.P. Alternately, if the clay containing the whale bone predates the Kent phase, the Chignecto phase could have occurred just before 13,500 B.P. (the assigned age of the Bantalor phase Hammond River maximum, which postdates the Chignecto phase). Radiometric dates of >43,000 (GSC-1680) on peat and 13,600 ± 220 B.P. (GSC-1222) on a mastadon bone from a sequence of silts in a karst depression near Hillsborough (Table 1) are difficult to interpret in light of the possibility that the karst development may in part precede the Chignecto phase.

Kent Phase

A strong easterly flow pattern, which postdates the Chignecto phase, has been documented in Kent County north of Moncton (Rampton and Paradis 1981b). Northeast-trending striae, the Pollett River flow pattern (Fig. 9) and southeast-trending striae on the northeast edge of the Caledonian Highlands have been assigned to the Kent phase. However, it is possible that these flow patterns could have been established during the late stages of the Chignecto phase or early stages of a later glacial phase (Bantalor phase). Correlation of glacial limits and flow patterns on the east and west sides of the Caledonian Highlands is difficult.

It is also assumed that the St. Martins flow pattern persisted during the Kent phase. The St. Martins flow pattern includes: (a) those south-trending striae and flutings west of Wood Lake that are slightly discordant with the south-southeast trend of the Caledonia flow pattern east of Wood Lake, and (b) those southeast- to southwest-trending striae north of Cape Spencer; but excludes those striae within the proposed limit of the Bantalor phase. The variable flow pattern direction near Cape Spencer is probably due to drawdowns of the glacier caused by topography, bathymetry and nearby ocean waters.

The Kent glacial limit is well marked north of Hillsdale where a sharp boundary marks terrain characterized by stream incision and mature dendritic drainage patterns, and terrain characterized by glacially-deranged drainage and glacial landforms. However, south of Hillsdale no obvious break in terrain character exists between terrain marked by the St. Martins flow pattern and terrain marked by the Caledonia flow pattern other than the flow patterns are slightly discordant and the Salmon River is deeply incised relative to streams to its west. The maximum extent of the Kent phase glaciers north of Sussex is difficult to determine as no obvious breaks in terrain character occur inside the Chignecto glacial limit. Areas of thick till southeast of Petitcodiac and south of Nixon may mark the Kent glacial limit.

The Pollett River, St. Martins flow patterns and other flow patterns on the New Brunswick Lowland (Rampton and Paradis 1981b) suggest a large ice centre or a series of ice caps stretching from a general area encompassing the drainage basin of the Gaspereau River (the Gaspereau ice centre) to the area west of Anagance (Fig. 9). Ice flowing from further northwest probably blocked any general southwest flow of ice from the Gaspereau ice cap.

During the interval when the Kent phase glaciers were near their maximum extent, streams draining east and north from the Caledonian central plateau would have been dammed and would have caused lakes to form in the dammed valleys as occasionally attested to by sand-veneered drift surfaces, absence of sharply delineated esker and morainal ridges, and accumulation of alluvium from the Caledonian central plateau along the glacier boundary. The general absence of extensive thick lacustrine deposits and meltwater channels suggests that little water was moving laterally between the ice and the Caledonian Highlands. Probably the peripheral lakes were ephemeral and drained subglacially as in the earlier Caledonia phase. The Walton Lake spillway and the channel between Mill Creek and the Salmon River south of Saddleback Mountain may have been reactivated

Table 1. Radiocarbon Dates and Descriptions

Sample Number	Map Site	Latitude (N)	Longitude (W)	Approx. Elev.	Details	Date
GSC-2167	A	45°13.7'	65°57.1'	20 m	*Portlandia arctica siliqua* (Reeve) shells from sand overlying clay and under 24 m of sand and gravel	13,000 ± 200
BGS-642	B	45°37.7'	64°50.5'	18 m*	Peat from base of bog overlying 17 m of glaciomarine sand and silt	7880 ± 120
BGS-668	C	45°47.5'	64°30.2'	15 m	Wood detritus in 0.5 m sandy unit overlain by 0.6 m of clayey silt sand and peat	1725 ± 90
BGS-669	C	45°47.5'	64°30.2'	15 m	Woody peat under 0.3 m sand and over 0.7 m sand and clayey silt	modern
BGS-666	D	45°57.2'	64°23.3'	55 m	Basal .02 m of 1.6m of peat overlying sand	9300 ± 140
GSC-2573	I	45°13.5'	65°57.2'	38 m	Valve of *Hiatella arctica* from 10 m below surface in sand pit (Lowden and Blake 1979)	14,400 ± 530
GSC-2640	I	45°13.5'	65°57.2'	28 m	Shell fragments of *Mya* sp. from 10 m below surface in sand pit (Lowden and Blake 1979)	12,800 ± 120
GSC-2544	II	45°37'	65°03.5'	380 m	Peat from 0.87 to 0.98 m below surface of bog (Lowden and Blake 1978)	830 ± 70
GSC-2524	II	45°37'	65°03.5'	380 m	Peat from 1.75 to 1.83 m below surface of bog (Lowden and Blake 1978)	6080 ± 80
GSC-1222	III	45°54'	64°39'	10 m (?)	Mastadon bone in silt sequence in sink hole (Schroeder and Arsenault 1978)	13,600 ± 220
GSC-1680	III	45°54'	64°39'	10 m (?)	Peat layer in silt sequence in sink hole (Schroeder and Arsenault 1978)	>43,000

Table 1. (Cont'd.)

Sample Number	Map Site	Latitude (N)	Longitude (W)	Approx. Elev.	Details	Date
GSC-1030	IV	45°51'	64°18'	-6.1 m*	Shells (*Mya arenaria*) 5.3 m below top of intertidal muds (Lowden and Blake 1970)	1760 ± 140
GSC-975	IV	45°51'	64°18'	-9.4 m*	Stump rooted in till and overlain by salt marsh sediments (Lowden and Blake 1970)	3520 ± 140
GSC-930	IV	45°51'	64°18'	-11.5 m*	Stump rooted in till and overlain by salt marsh sediments (Lowden and Blake 1970)	4040 ± 130
GSC-1089	IV	45°51'	64°18'	-10.5 m*	White pine stump rooted in till and overlain by salt marsh sediments (Lowden and Blake 1970)	4120 ± 130
I-3241	V	45°53'	64°21'	-7.3 m*	Wood in salt marsh sediment sequence (MacNeill 1969)	3290 ± 100
I-3238	V	45°53'	64°21'	-7.3 m*	Wood in salt marsh sediment sequence (MacNeill 1969)	3430 ± 100
GSC-602	VI	45°57'	64°20'	-	Birch stump rooted in till and under one foot of salt marsh mud (Lowden and Blake 1970)	640 ± 130
I-4313	VI	45°57'?	64°20'?	-6.1 m*	Wood stump from 6.1 m below surface of marsh sediments (Welstead 1971)	1875 ± 95
I-4312	VII	45°57'?	64°15'?	-2.1 m*	Organic material 2.1 m below surface of marsh sediments (Welstead 1971)	935 ± 95

*Indicates approximate level relative to high tide.

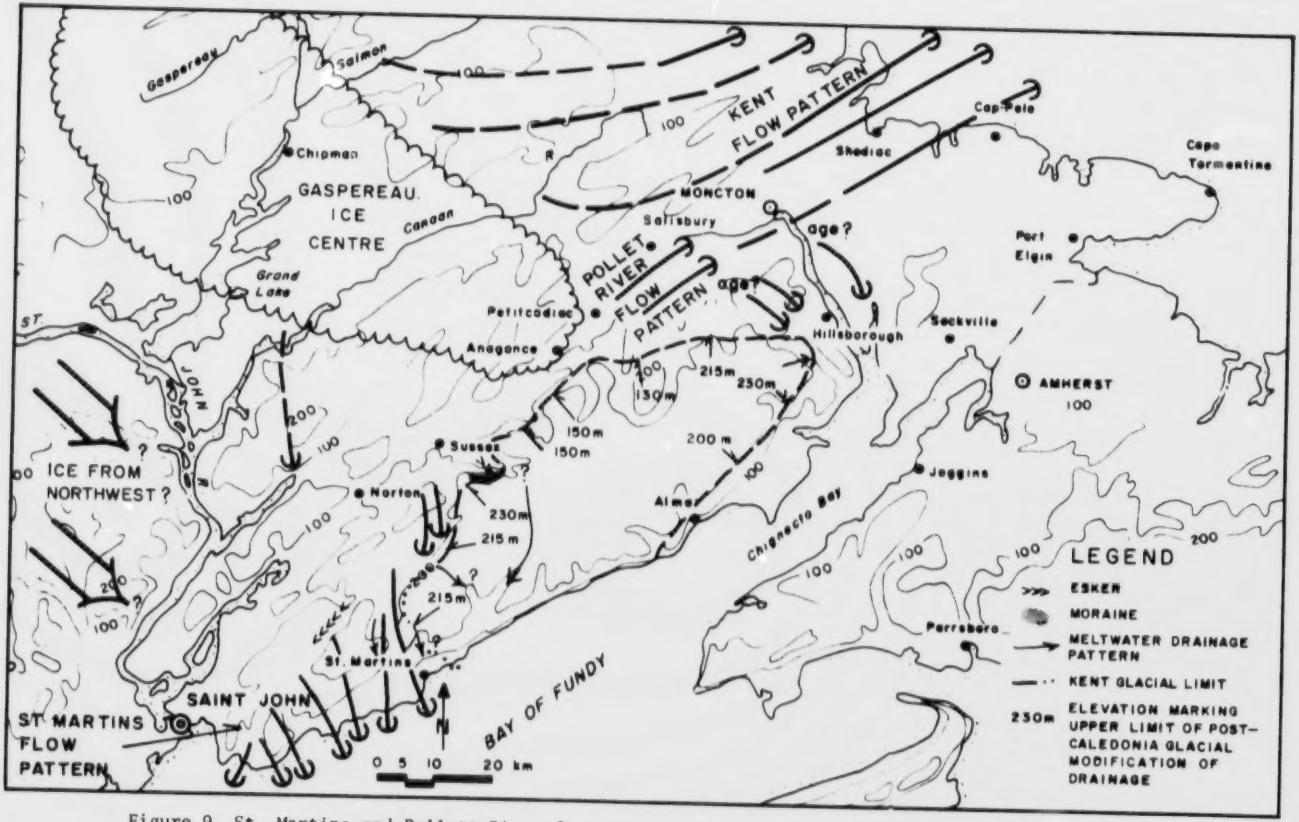

Figure 9. St. Martins and Pollett River flow patterns and glacial features of Kent phase.

during the Kent glacial maximum. Certainly as the ice waned and evacuated the Hammond River valley north of Hanford Brook, meltwater moved south along this part of the Hammond River valley from the wind gap between Ward Creek and the Hammond River.

During the interval when the land between St. Martins and Loch Lomond was in the early stages of deglaciation, water was moving subglacially across the drainage divide between the Hammond River and Bay of Fundy, and eskers and kame complexes were formed on the uphill side and crest of the drainage divide. A major esker system near Barnsville formed where meltwater moving down the Hammond River and Salt Spring Brook drainage systems was forced through the drainage divide into the Loch Lomond drainage system. As the ice retreated from the area bordering the Bay of Fundy, outwash deposits and glaciomarine deltas were deposited in the area encompassed by Black River, Shanklin and St. Martins (Finamore and Lohse 1977).

The Kent phase is believed to have occurred prior to 18,500 B.P. as areas covered during this phase near the Northumberland Strait were deglaciated by this time (Rampton and Paradis, 1981b). Dates of 12,800 + 120 B.P. (GSC-2640), 13,000 + 200 B.P. (GSC-3167) and 14,400 + 530 B.P. (GSC-2573) were obtained on shells collected from a delta north of the mouth of the Mispec River (Table 1). However, these dates are believed to relate to postglacial marine submergence, or an interval when meltwater was diverted down the Mispec River by ice of a later phase (the Bantalor phase). Deglaciation of the area covered by the St. Martin flow pattern could have occurred before or after the Kent phase. This depends on the validity of correlation of the St. Martin flow pattern with the Pollett River and other flow patterns to the north of Moncton. Deglaciation appears to have occurred well before maximum marine submergence as most glaciomarine deltas are between 22 m (75 ft) and 45 m (150 ft) elevation along the Fundy shoreline between Quaco Bay and Black River (Finamore and Lohse 1977), whereas maximum marine submergence appears to be over 75 m (250 ft) elevation in the same area.

Bantalor Phase

During deglaciation of the New Brunswick Lowland northwest of Moncton, a ridge of glacier ice was established that extended from the upper reaches of the Gaspereau River through to Anagance (Rampton and Paradis 1981b) and led to northeast- and south-moving ice (Fig. 10). The interval during which this ice centre and flow patterns persisted has been termed the Bantalor phase.

The eastern margin of ice during the Bantalor phase was probably east of Moncton, but its location is difficult to identify; perhaps because it was calving into the ocean. A belt of moraine east of Stoney Creek and near Riverview Heights south of Moncton (Rampton and Paradis 1981b) are possible terminal positions. As indicated previously, the Pollett River flow pattern may have originated during the Kent phase, but probably persisted during the early part of the Bantalor phase.

The primary glacial flow, the Norton flow pattern, during the Bantalor phase is defined by numerous south-southeast-trending striae in the area north of a line extending from Lakewood through Jeffries Corner (Fig. 10). The Norton flow pattern suggests that the primary source of ice was from the Gaspereau ice centre but that coalescent glacier ice to the west prevented ice from the Gaspereau ice centre from flowing to the southwest.

During the late stages of the Bantalor phase ice retreated from its maximum extent on Figure 10, to northwest of Grand Lake in the northwest corner of the area. At its maximum extent, ice of the Bantalor phase is envisaged to have constructed the large complex of ablation drift and ice-contact deposits at the southwest end of Loch Lomond and in the Little River valley (Melvin's (1966) Little River 'stratified drift' sequence), and possibly diverted meltwater down the Mispec River (Melvin's (1966) Mispec River 'stratified drift' sequence). Other features associated with the Bantalor maximum are morainal topography along hill slopes east of Hammond, an ablation complex near Hanford Brook, ribbed and rolling moraine between Upham and Cassidy, and a major complex of outwash and ablation drift along the southeastern edge of the Kennebecasis River valley

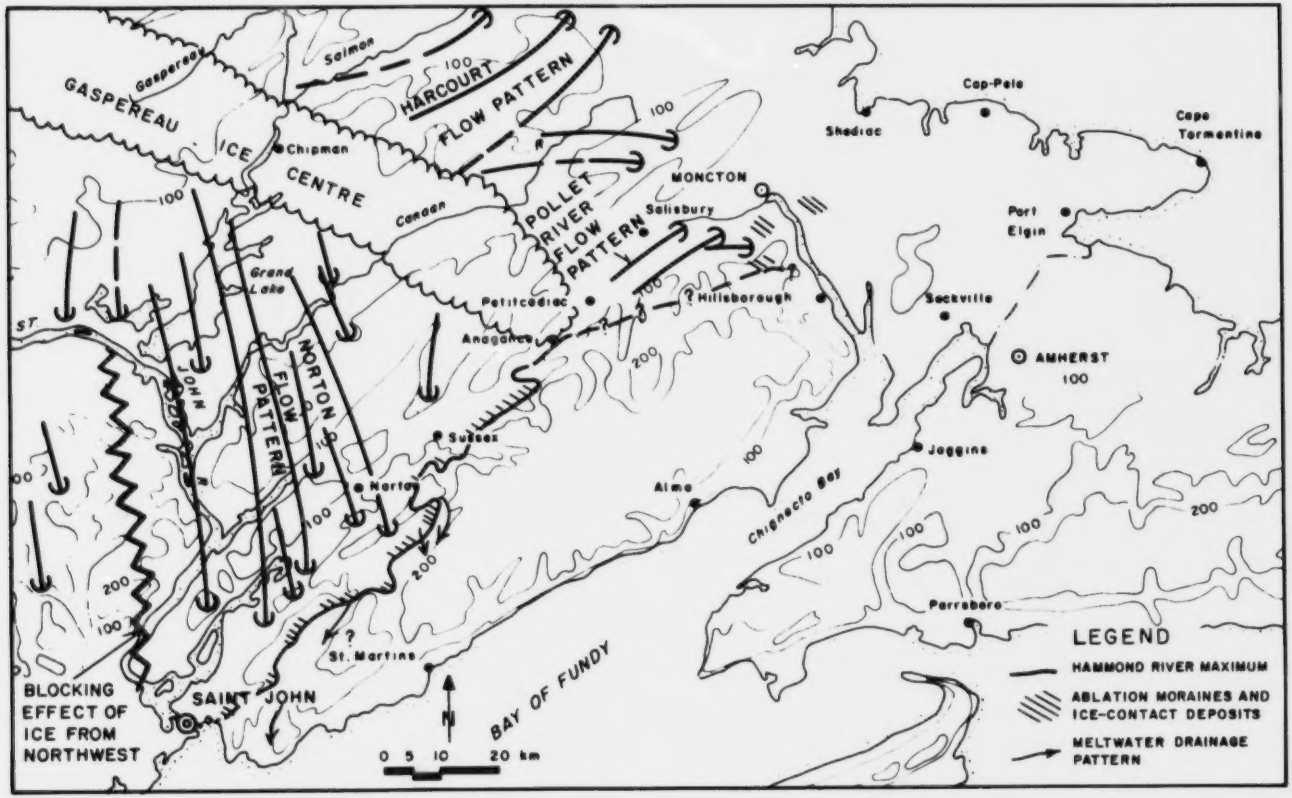

Figure 10. Glacial features of Bantalor phase.

northeast of Sussex (Brinsmead 1977). It is possible that some ice-contact gravels at the latter locality were deposited as alluvium on the glacier ice by the Kennebecasis River flowing from the glacier-free Caledonian Highlands. Deltaic deposits (Brinsmead 1977) in this area also may have been deposited by streams flowing from glacier-free areas into lakes along the edge of the glacier. During the maximum extent of the Bantalor phase, defined here as the Hammond River maximum, meltwater probably moved into the upper reaches of Hammond River through gaps in the bedrock ridges east of Cassidy Lake, and at Hillsdale. Drainage from this area may have reoccupied the glacial spillway at Barnsville, but subglacial drainage along the present westward course of the Hammond River is more probable. At Ben Lomond meltwater was diverted down the Mispec River. The Hammond River maxima was probably synchronous with Gadd's (1973) Pennfield phase.

Tracing of possible still-stands of the retreating Bantalor phase ice is hampered by a lack of knowledge of the influence exerted by the northeast-trending bedrock ridges in the ridge and valley area and by marine submergence. Thinning of glacier ice would have had the effect of isolating blocks of ice in broad valleys (Brinsmead and Finamore 1977), and their flow would then have been controlled by local topography and the effects of ponded or marine water. In spite of this, belts of rolling ablation moraine and ice-contact deposits define ice margins of more than local significance that appear to correspond with Gadd's (1973) ice margin positions or phases (Fig. 11).

Ice-contact deposits and outwash near Bradley Lake (Melvin's (1966) Bradley Lake-Carpenter Pond Sequence), belts of ablation moraine in the Salt Spring Brook valley and near Southfield (Brinsmead and Finamore 1977) and part of the morainal complexes east and north of Sussex define a pause in the ice marginal position that is termed the Salt springs retreatal phase. The complex of outwash, ice-contact deposits, and rolling morainal deposits that occupy the valley of Smith Creek, scattered kame terraces and morainal deposits along the middle reaches of the Kennebecasis River valley and an area of moraine near McManus Lake, define an ice marginal position that is termed the Smith Creek retreatal phase. This phase can probably be equated to Gadd's (1973) Bethel-Pocologan phase. The esker/outwash complex north of Hampton was probably developed during the Smith Creek retreatal phase. During the Salt Springs or Smith Creek retreatal phases anomalous striations recorded near Lower Millstream and esker segments west of Sussex (Brinsmead and Finamore 1977) indicate that ice may have flowed parallel to the main valleys once the glacier ice thinned to a critical thickness that prevented the regional flow from persisting. Scattered occurrences of silty sand, silt and clay near Norton (Brinsmead and Finamore 1977) and near Sussex indicate that these glaciers were either calving into glacially dammed lakes or estuarine waters.

Some evidence of glacier retreat and readvance has been noted within the area covered by ice of the Salt Springs and Smith Creek retreatal phases. Near Penobsquis clays were noted that were capped by till, near Five Points Brinsmead (1977) noted till over glaciofluvial deposits and compact till layers were noted in the outwash sequence east of Knightville along Smith Creek.

A major complex of rolling ablation moraine and ice-contact deposits along the Millstream River and Belleisle Creek valleys defines a pause in ice marginal retreat (Brinsmead and Finamore 1977) that is termed the Belleisle retreatal phase. This phase would appear to correlate with Gadd's (1973) McDougall Lake phase. During the latter part of the Belleisle retreatal phase, it would appear that meltwater channels crossing the bedrock ridge north of the Belleisle Creek/Millstream River valley debouched outwash upon glacier ice stagnating in this valley.

Following the Belleisle retreatal phase, no well defined morainal or ice-contact system was constructed in the area. An apparent area of thick till adjacent to Washademoak Lake is perhaps an eastward extension of Gadd's (1973) Shin Creek phase. A ridged, rolling belt of moraine north of Youngs Cove Road appears to be part of an ice marginal position defined by the New Canaan moraine to the east (Rampton and Paradis 1981b) and Gadd's (1973) Oromocto-Gagetown phase to the west.

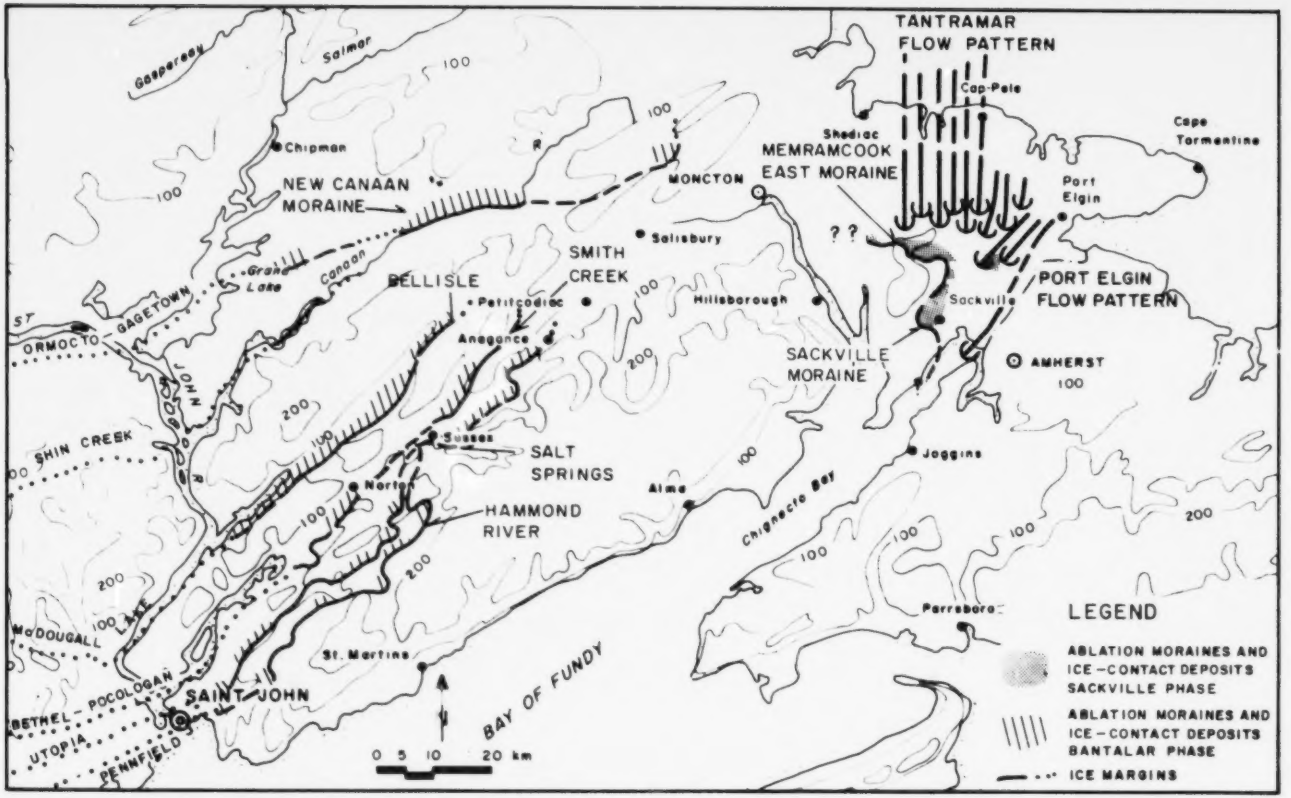

Figure 11. Glacial features of Sackville phase and ice marginal features of Bantalor retreatal phases.

Glacier ice appears to have retreated from the Hammond River Maximum by 13,500 B.P. Radiocarbon dates of 12,800 ± 120 B.P. (GSC-2640), 13,000 ± 200 B.P. (GSC-3167) and 14,400 ± 530 B.P. (GSC-2573) on shells in sand near the mouth of the Mispec River may relate to meltwater being diverted down the Mispec River during the Hammond River maximum. Mott (personal communication 1981) obtained a date of 13,100 ± 280 (GSC-2787) from the basal organic sediments in Cassidy Lake, and Gadd (1973) has numerous dates between 13,000 and 13,300 B.P. on marine submergence, which postdated the Pennfield stage. Older basal dates obtained by Gadd in a kettle on outwash of his Pennfield stage are believed to be invalid because of the hardwater effect (Mott 1975). The oldest dated moraine in Maine is believed to have been constructed around 13,500 B.P. (Borns 1973).

The Salt Springs and Smith Creek retreatal phases contain evidence of readvance within their boundaries. Perhaps these readvances are the equivalent of the Kennebunk glacial advance dated at 13,200 B.P. or the Pineo Ridge readvance dated at 12,700 B.P. in Maine (Borns 1973). The Gaspereau phase probably ended well before 12,000 B.P. as deglaciation of most of the Saint John River valley was completed by 11,200 B.P. (Rampton and Paradis 1981a).

Sackville Phase

A belt of hummocky and rolling moraine at Sackville and ribbed moraine north of Joe Brook (Fig. 11) define an ice marginal position for ice flowing south from Northumberland Strait or Prince Edward Island (Tantramar flow pattern) and ice flowing southwest in the Baie Verte area (Port Elgin flow pattern). During the Sackville phase meltwater appears to have moved parallel to the western edge of the glacier down Joe Brook and south of Ogden Mills. The ice-contact and outwash deposits in the outlet at Frosty Hollow was probably deposited subaqueously, either in a lake dammed by a glacier to the east or in an area flooded by marine waters.

The Sackville phase may have been more complex than outlined here. Ice flow of the Tantramar and Port Elgin flow patterns do not appear to have always been synchronous; and limits of marine submergence at Sackville and northeast of Midgic appear to be discordantly low relative to those west of Sackville and at the head of the Memramcook River. These apparent discordant marine limits might be explained by a pause in retreat which allowed some marine emergence to occur between retreat of the ice from the Memramcook East and the western part of the Sackville moraine, and its retreat from the eastern part of the Sackville moraine.

The Sackville phase appears to postdate the Chignecto and Kent phases. Evidence for this is the lower indications of marine submergence on drift of the Sackville phase than on drift of the Chignecto phase (Fig. 12); and superposition of striae assigned to the Sackville phase on those assigned to the Kent phase (Rampton and Paradis 1981b). The Sackville phase was probably in part contemporaneous with the Bantalor phase, and may correlate with the early part of Prest's (1973) Prince Edward Island ice-cap phase of glaciation. The Sackville phase probably predates 12,500 B.P. because the western part of Prince Edward Island was deglaciated by this time (Prest 1973).

Late Glacial and Postglacial Events

Late Wisconsin and Early Postglacial Submergence

As indicated previously, the absence of sharp-crested moraines and eskers and kettles at low elevations suggest that ice may have been calving directly into marine and estuarine waters or that these lower elevations were submerged following deglaciation. Generally most evidence in southern New Brunswick supports the concept of maximum marine submergence postdating deglaciation (Gadd 1973, Rampton and Paradis 1981a,b).

The presence and distribution of glaciomarine deltas, marine fossils, marine (?) clays, sand veneers, upper limits of sand veneers, and lower limits

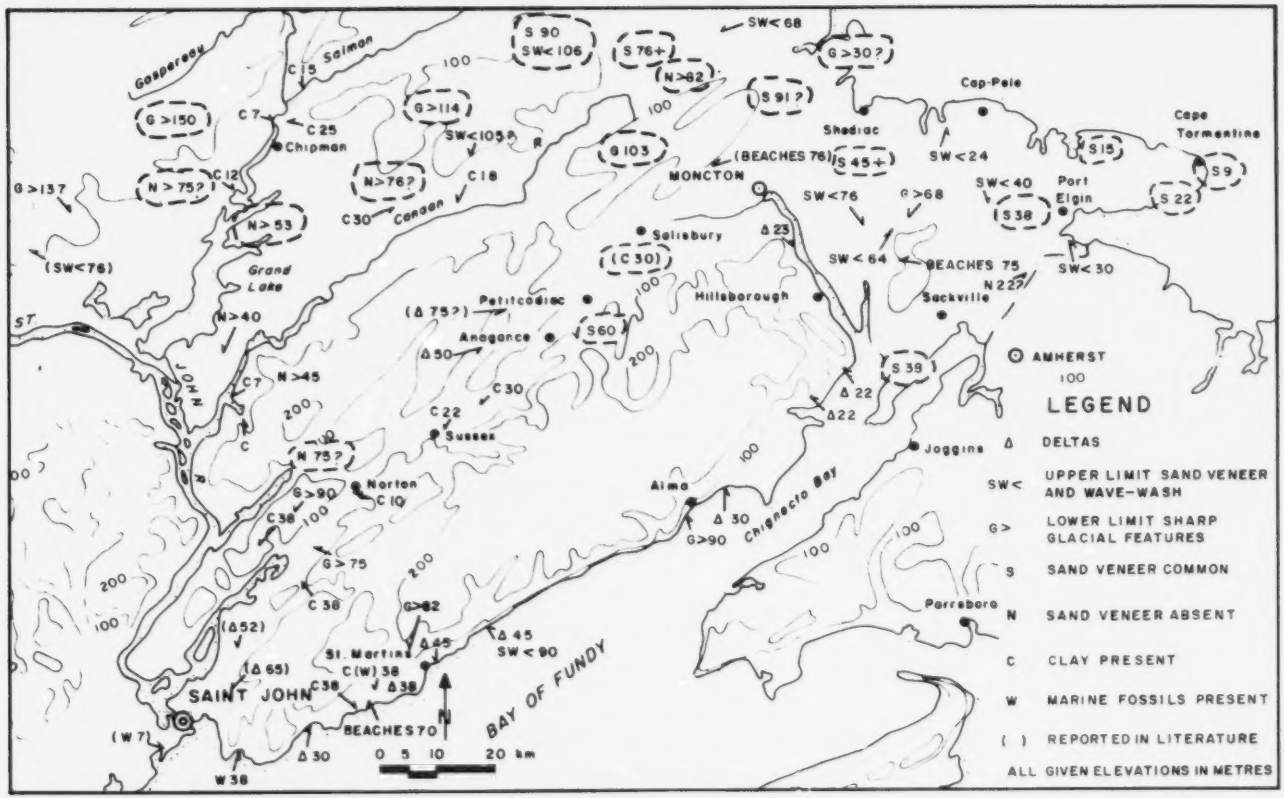

Figure 12. Indicators of marine submergence. The elevations of phenomena suggesting marine submergence and a limit of marine submergence (marine limit) are plotted.

of sharply-defined kettle holes, eskers, and morainic ridges have been plotted to derive possible marine limits throughout eastern New Brunswick (*cf.* Fig. 12). These marine limits do not appear to have been synchronous, possibly because of the complex relation of rates and timing of deglaciation, rising sea level and crustal rebound. A few discordancies in elevations of marine limit may be explained by the existance of residual ice caps, especially the Escuminac and Northumberland Strait ice centres. Their presence would have delayed submergence of terrain covered by them until after isostatic rebound had caused regression to commence in the general area.

Numerous marine deltas between Black River and Hillsborough on the Fundy coast having surface elevations between 22 m (75 ft) and 45 m (150 ft) indicate that sea level was near these elevations during retreat from this area of Chignecto and, probably, Kent phase glaciers (these events are presumed to predate 18,500 B.P.). Beaches south of Fairfield, the maximum height of sand at Salmon River, and the limit of hummocky ice-contact deposits near Alma are between 75 m (250 ft) and 90 m (300 ft) in elevation (Fig. 12) and indicate that the elevation of maximum marine submergence was higher than noted elevations of marine deltas along the Fundy coast. Projected isobases on the levels of maximum marine submergence are more or less parallel to the coast, and would appear to concur with levels of marine submergence (Prest *et al.* 1968) and isobases (Grant 1980a) along parts of the southern Fundy coast that were probably not glaciated during the latter part of the Wisconsin. Beaches (?), near 75 m (250 ft) elevation, along the Trans Canada Highway east of College Bridge (Fig. 12) also appear to confirm these isobases, but sand-veneered drift at higher elevations confuses the identification of maximum levels of marine submergence in that area. Marine limit in areas covered by Chignecto and Kent phase glaciers may locally be higher than in nearby areas covered by Bantalor and Sackville phase glaciers (if significant rebound and marine regression occurred between the Kent and Bantalor phases).

The maximum level of marine submergence along the northern and western edge of the Caledonian Highlands is extremely difficult to determine because of the presence of a relatively continuous sand veneer and many pockets of silt and clay, which are not limited to specific elevations. The sand veneer is believed to be due to wave-washing, but much of it is probably related to proglacial lakes, rather than marine submergence.

In areas covered by the Bantalor and Sackville phases, limits of marine submergence are only locally well defined. Melvin (1966) has indicated that the limit of submergence is at an elevation of about 67 m (220 ft) in the area west of Ben Lomond. This is based on the distribution of glaciomarine (?) deltas, marine sediments and wave-washed surfaces. Gadd (1973) believes that submergence never exceeded 73 m (240 ft) elevation near Pennfield to the west. In the Kennebecasis, Petitcodiac and Millstream valleys marine submergence is probably limited to elevations below 75 m (250 ft) or 90 m (300 ft) as defined by glaciomarine deltas (Brinsmead 1977) and the lower limit of sharp morainic and esker ridges. Near Grand Lake, the upper elevation of washed drift appears to be between 45 m (150 ft) and 60 m (200 ft). This appears anomalously low compared to apparent levels of submergence to the north and west, but may have been caused by the persistance of late ice near Minto (Rampton and Paradis 1981a,b). Apparent maximum levels of submergence of between 15 m (50 ft) and 30 m (100 ft) elevation in areas covered by the Sackville phase near Sackville and northeast of Sackville may also be due to the relatively young age of the Sackville phase.

The area of apparent marine submergence and wave-washing as indicated on Map 81-003, in Fig. 12, and in adjacent areas (Rampton and Paradis 1981a,b) clearly includes areas higher than the limit of submergence indicated by other recent publications (*cf.* Prest *et al.* 1968, Grant 1980a). However, their limits are based in part on extrapolation of evidence from adjacent areas, and a simplified model of deglaciation. Evidence presented herein for a complex history of glaciation and deglaciation support discordancy in levels of marine submergence from area to area, and higher limits of submergence in much of the area. Levels of maximum submergence in Maine (Leavitt and Perkins 1935) also support high levels of marine submergence in adjacent areas to the northwest (Rampton and Paradis 1981a,b).

Postglacial Emergence

Little direct evidence of the timing of emergence in southeastern New Brunswick is available. However, dates of 9300 ± 140 B.P. (BGS-666) from just west of Sackville and 7880 ± 120 B.P. (BGS-642) from near Waterside (Table 1) on peats overlying marine sediments and glacial drift give minimal ages for emergence of these areas. BGS-668 (1725 ± 90) and BGS-669 (modern) were originally thought to date emergence near Johnson Mills, but must be related to recent alluvial activity at the collection site.

In adjacent Maine, Bloom (1963) suggests that emergence of the land surface was completed there by 7500 B.P., whereas Stuiver and Borns (1975) believe it was completed by 12,100 B.P. In the environs of Prince Edward Island it was completed by 7000 B.P. (Kranck 1972, Prest 1973).

Along the eastern coast of New Brunswick, salt marsh peat and oysters collected from below sea level and dated in the interval between 7500 and 3000 B.P. (Kranck 1972, Prest 1972) indicate that relative sea level was well below its present level during that interval. Stumps nearly 12 m below high tide level at Fort Beausejour dated at 4010 ± 130 B.P. (GSC-930) indicate that sea level was probably locally below this level at some time before 4000 B.P.

Late Postglacial Submergence

During the last 4000 years the land along the Bay of Fundy has been submerging (cf. Grant 1970) as indicated locally by wood between 6.7 and 7.6 m below the present level of the Tantramar Marsh and dated at 3430 ± 110 B.P. (I-3238) and 3290 ± 100 B.P. (I-3241), and stumps buried in intertidal mud from levels of 9.4 to 11.5 m below high tide level near Fort Beausejour and dated at 3520 ± 140 B.P. (GSC-975), 4040 ± 130 B.P. (GSC-930) and 4120 ± 130 B.P. (GSC-1089). *Mya* shells dated at 1790 ± 130 B.P. (GSC-1030) in mud at 6.1 m below high tide level indicate that sea level was above that level by 1800 B.P. Grant (1970) has calculated that sea level is rising in the Bay of Fundy at a rate of 0.30 m/100 years (one ft/100 years). Extensive areas of intertidal mud and salt marsh along Chignecto Bay are probably due to the rising sea level and high amplitude of tides in the Bay of Fundy.

Terrestrial and Alluvial Events

During the postglacial surfaces uncovered by glacier ice and water have been subjected to weathering and soil formation (cf. Aalund and Wicklund 1949). Bogs and peatlands have developed in many depressions and swales characterized by poor drainage or impervious soils; a development that appears to have intensified around 8000 B.P. to 8500 B.P. (Korpijaakko 1976) and continues at present. Locally basal organics in areas covered by peat have been dated (Table 1) at 9300 ± 140 B.P. (BGS-666) and 7880 ± 120 (BGS-642). In Fundy National Park peat at about 1.8 m below the surface was dated at 6080 ± 80 B.P. (GSC-2524) and at 0.9 m at 830 ± 70 B.P. (GSC-2544) indicating variable rates of peat accumulation.

Karst topography has developed in a number of areas around the flanks of the Caledonian Highlands where gypsum and anhydrite occur. Most of this Karst topography is believed to be postglacial in age, but one karst sink hole contains peat dated at >43,000 B.P. (GSC-1680) and a mastadon bone dated at 13,600 ± 220 (GSC-1222) in a sequence of silts overlying till and bedrock near Hillsborough (Schroeder and Arsenault 1978).

Most streams have relatively narrow terraces and floodplains, although locally broad terraces and floodplains have developed. Whereas most stream courses are presently downcutting to adjust to lower sea levels that occurred during postglacial emergence, floodplain development and alluviation along the lower courses of streams is an adjustment to the more recent submergence, which is continuing in the area.

REFERENCES

Aalund, H. and Wicklund, R.E. 1949: Soil survey report of southeastern New Brunswick; Third rept. of the New Brunswick Soil Survey, N.B. Dept. Agric., Fredericton.

Bloom, A.L. 1963: Late-Pleistocene fluctuations of sea level and postglacial crustal rebound in coastal Maine; Am. Jour Sci., V. 261, pp. 862-879.

Borns, H.W., Jr. 1963: Preliminary report on the age and distribution of the late Pleistocene ice in north central Maine; Am. Jour. Sci., V. 261, pp. 738-740.

Bostock, H.S. 1970: Physiographic regions of Canada; Geol. Surv. Can., Map 1254A.

Brinsmead, R.A. 1974a: Surficial geology and granular resources of Petitcodiac East (21 H/14 E); N.B. Dept. Nat. Res., Min. Res. Br., Plate 74-121.

_____ 1974b: Surficial geology and granular resources of Hillsborough West (21 H/15 W); N.B. Dept. Nat. Res., Min. Res. Br., Plate 74-123.

_____ 1974c: Surficial geology and granular resources of Salisbury East (21 I/E); N.B. Dept. Nat. Res., Min. Res. Br., Plate 74-127.

_____ 1974d: Surficial geology and granular resources of Salisbury West (21 I/3W); N.B. Dept. Nat. Res., Min. Res. Br., Plate 74-128.

_____ 1977: Aggregate resources of Petitcodiac West map area (21 H/14 W); N.B. Dept. Nat. Res., Min. Res. Br., Topical Rept. 77-7.

_____ 1978: Granular aggregate resources, Sevogle map area (21 P/4); N.B. Dept. Nat. Res., Min. Res. Br., Open File 78-10.

Brinsmead, R.A. and Finamore, P.F. 1977: Aggregate resources of the Sussex map-area (21 H/12) and part of the Hampstead map area (21 G/9E) including surficial geology of Sussex East (21 H/12 E); N.B. Dept. Nat. Res., Min. Res. Br., Topical Rept. 77-8.

Carroll, D.J.J. 1977: List of publications available from Mineral Resources Branch, Dept. of Natural Resources; N.B. Dept. of Nat. Res., Informat. Circ. 76-1.

Chalmers, R. 1884: Report on the surface geology of western New Brunswick with special reference to the area included in York and Carleton Counties; Geol. Surv. Can., Rept. of Progress 1882-84, pt. GG.

_____ 1886: Preliminary report on the surface geology of New Brunswick; Geol. Surv. Can., Summ. Rept. 1884 and 1885, v. I, pt. GG.

_____ 1887: Report to accompany quarter-sheet maps 3 S.E. and 3 S.W. surface geology; northern New Brunswick and southeastern Quebec; Geol. Surv. Can., Ann. Rept. 1886, v. II pt. M.

_____ 1888: Report on the surface geology of northeastern New Brunswick; Geol. Surv. Can., Summ. Rept. 1887 and 1888, v. III, Pt. N.

_____ 1890: Report on the surface geology of southern New Brunswick; Geol. Surv. Can., Summ. Rept. 1889, v. IV, pt. N.

_____ 1895: Report on the surface geology of eastern New Brunswick, northwestern Nova Scotia, and a portion of Prince Edward Island: Geol. Surv. Can., Ann. Rept. 1894, v. VII, Rept. M.

_____ 1901: Work on the surface geology of northwestern New Brunswick chiefly in the area of the Grand Falls sheet; Geol. Surv. Can., Ann. Rept. 1900, v. XIII, pt. A., pp. 151-161.

_____ 1902: Report on the surface geology shown on the Fredericton and Andover quarter-sheet maps, New Brunswick; Geol. Surv. Can., Ann. Rept. 1899, v. XII, Rept. M, 36 p.

Finamore, P.F. in press: Granular aggregate resources of Millville (21 J/3) and Woodstock (21 J/4); N.B. Dept. Nat. Res., Min. Res. Br.

Finamore, P.F. and Lohse, H. 1977: Granular resources in the Loch Lomond map area (21 H/5); N.B. Dept. Nat. Res., Min. Res. Br., Topical Rept. 77-9.

Gadd, N.R. 1973: Quaternary geology of Southwest New Brunswick, with particular reference to Fredericton Area; Geol. Surv. Can., Paper 71-34.

Gauthier, R.C. 1978: Quelques interprétations de l'inventaire des dépôts de surface, péninsule nord-est du Noveau Brunswick; Geol. Surv. Can., Paper 78-1A, pp. 409-412.

Gauthier, R.C. 1979: Aspects of the glacial history of the north-central Highlands of New Brunswick; Geol. Surv. Can., Paper 79-1B, pp. 371-377.

_____ 1980a: Existence of a central New Brunswick ice cap based on evidence of northwestward-moving ice in the Edmundston area, New Brunswick; Geol. Surv. Can., Paper 80-1A, pp. 377-378.

_____ 1980b: Decomposed granite, Big Bald Mountain area, New Brunswick; Geol. Surv. Can., Paper 80-1B, pp. 277-282.

Gauthier, R.C. and Thibault, C. 1980: Deglaciation of the Edmundston area and reappraisal of Glacial Lake Madawaska interpretation; pp. 188-190 *in* a guidebook to the geology of northeastern Maine and neighbouring New Brunswick, *eds.* D.C. Roy and R.S. Naylor, 72nd Ann. Meeting N.E.I.G.C.

Grant, D.R. 1970: Recent coastal submergence of the Maritime Provinces, Canada; Can. Jour. Earth Sci., v. 7, pp. 676 - 689.

_____ 1977: Glacial style and ice limits, the Quaternary stratigraphic record, and changes of land and ocean level in the Atlantic Provinces, Canada; Géographie, physique et Quaternaire, v. 31, pp. 247-260.

_____ 1980a: Quaternary sea level change in Atlantic Canada as an indication of crustal delevelling; pp. 201-214 *in* Earth rheology, isostasy and eustasy, *ed.* Nils-Axel Mörner, John Wiley and Sons Ltd., New York.

_____ 1980b: Quaternary stratigraphy of southwestern Nova Scotia; glacial events and sea level changes; Geol. Ass. of Can. Ann. Meeting, Halifax 1980, Field Trip 9 Guidebook.

Greiner, H. 1974: Geomorphology of the Fundy National Park, New Brunswick; Maritime sed., v. 10, pp. 36-51.

Hamilton, J.B. 1973: Generalized surficial geology, illustrating the distribution of granular resources in New Brunswick; N.B. Dept. Nat. Res., Min. Res. Br., Plate 73-14.

Kiewiet de Jonge, E.J.C. 1951: Glacial water levels in the Saint John River valley; Ph.D., thesis, Clark University, Worrester, Massachusetts.

Kingston, P.W. 1974a: Surficial geology and granular resources of Blackville East (21 I/12 E); N.B. Dept. Nat. Res., Min. Res. Br., Plate 74-130.

_____ 1974b: Granular aggregate resources of the Moncton-centred region, summary; N.B. Dept. Nat. Res., Min. Res. Br., Topical Rept. 74-7.

Korpijaakko, M.L. 1976: On the origin and geobotanical development of muskeg in New Brunswick; Ph.D. thesis, Univ. of N.B., Fredericton, N.B.

Kranck, K. 1972: Geomorphological development and post-Pleistocene sea level changes, Northumberland Strait, Maritime Provinces; Can. Jour. Earth Sci., v. 9, pp. 835-844.

Leavitt, H.W. and Perkins, E.H. 1935: Glacial geology of Maine; Maine Technology Exp. Sta., Bull. no. 3.

Lee, H.A. 1955: Surficial geology of Edmundston, Madawaska and Temiscouata Counties; New Brunswick and Quebec; Geol. Surv. Can., Paper 55-15.

_____ 1957: Surficial geology of Fredericton, York and Sunbury Counties, New Brunswick; Geol. Surv. Can., Paper 56-2.

_____ 1959a: Surficial geology of Grand Falls, Madawaska and Victoria Counties, New Brunswick; Geol. Surv. Can., Map 24-1959.

_____ 1959b: Surficial geology, Aroostock, Victoria County, New Brunswick; Geol. Surv. Can., Map 34-1959.

_____ 1962: Surficial Geology of Canterbury, Woodstock, Florenceville and Andover Map Areas, York, Carleton and Victoria Counties, New Brunswick; Geol. Surv. Can., Prelim. Paper 62-12.

Lohse, H. 1974a: Surficial geology and granular resources of Hillsborough East (21 H/15 E); N.B. Dept. Nat. Res., Min. Res. Br., Plate 74-122.

_____ 1974b: Surficial geology and granular resources of Moncton East (21 I/2 E); N.B. Dept. Nat. Res., Min. Res. Br., Plate 74-125.

_____ 1974c: Surficial geology and granular resources of Moncton West (21 I/2 W); N.B. Dept. Nat. Res., Min. Res. Br., Plate 74-126.

_____ 1975: Surficial geology and granular resources of part of Port Elgin West (21 I/W); N.B. Dept. Nat. Res., Min. Res. Br., Plate 74-124.

Lowden, J.A., and Blake, W.Jr, 1970: Geological Survey of Canada radiocarbon dates IX; Geol. Surv. Can., Paper 70-2B.

_____ 1978: Geological Survey of Canada radiocarbon dates XVIII; Geol. Surv. Can., Paper 78-7.

Lowden, J.A., and Blake, W. Jr. 1979: Geological Survey of Canada radiocarbon dates XIX; Geol. Surv. Can., Paper 79-7.

MacNeill, R.H. 1969: Dating some Quaternary changes in sedimentation in the Tantramar marsh of New Brunswick; Maritime Sed. v. 5, pp. 1-2.

Matthew, G.F. 1879: Report on the superficial geology of southern New Brunswick 1878; Geol. Surv. Can. Rept. of Progress for 1877-78.

Melvin, R.L. 1966: The surficial geology of Ben Lomond area, Saint John and kings Counties, New Brunswick; M.Sc. Thesis, Univ. of N.B., Fredericton, N.B.

Mott, R.J. 1975: Palynological studies of lake sediment profiles from south-western New Brunswick; Can. Jour. Earth Sci., v. 12, pp. 273-288.

New Brunswick Department of Natural Resources 1979: Distribution of peatlands; Map.

Poole, W.H., Sanford, B.V., Williams, H., and Kelly, D.G. 1970: Geology of the southeastern Canada; pp. 227-304 *in* Geology and economic minerals of Canada, *ed.* R.J.W. Douglas, Geol. Surv. Can. Econ. Rept. No. 1.

Potter, R.R., Hamilton, J.B., and Davies, J.L. 1979: Geological amp of New Brunswick; N.B. Dept. Nat. Res., Map N.R.-1.

Prest, V.K. 1973: Surficial deposits of Prince Edward Island; Geol. Surv. Can., Map 1366A.

Prest, V.K., Grant, D.R., Rampton, V.N. 1968: Glacial map of Canada; Geol. Surv. Can., Map 1253A.

Prest, V.K., Grant, D.R., Borns, H.W., Jr., Brookes, I.A., MacNeill, R.H., Ogden, J.G., III, Jones, J.R., Lin, C.L., Hennigar, T.W., and Parsons, M.L. 1972: Quaternary geology, geomorphology and hydrogeology of the Atlantic Provinces; 24th Internat. Geol. Congress, Guidebook A/C-61.

Rampton, V.N. and Paradis, S., 1981a: Quaternary geology of the Woodstock map area (21 J), New Brunswick; N.B. Dept. Nat. Res., Map Rept. 81-1.

---------- 1981b: Quaternary geology of the Moncton map area (21 I), New Brunswick; N.B. Dept. Nat. Res., Map Rept. 81-2.

Ruitenberg, A.A., McCutcheon, S.R., and Venugopal, D.V. 1976: Recent gravity sliding and coastal erosion, Devil's Half Acre, Fundy Park, New Brunswick: geological explanation of an old legend; Geoscience Can., v. 3, pp. 237-239.

Ruitenberg, A.A., Giles, P.S., Venugopal, D.V., Buttimer, S.M., McCutcheon, S.W., and Chandra, J. 1979: Geology and mineral deposits, Caledonia area; N.B. Dept. Nat. Res., Min. Res. Br., Memoir 1.

Schroeder, J., and Arseneault, S. 1978: Discussion d'un karst dans le gypse d'Hillsborough, Nouveau-Brunswick; Geogr., phys., Quat., v. 32, pp. 249-261.

Stobbe, P.C. 1940: Soil survey of the Fredericton-Gagetown area, New Brunswick; First Rept. of the New Brunswick Soil Survey, N.B. Dept. Agric., Fredericton.

Stuiver, M., and Borns, H.W., Jr., 1975: Late Quaternary marine invasion in Maine: its chronology and associated crustal movements; Geol. Soc. Am. Bull., v. 86, pp. 99-104.

Thibault, J. 1978: Granular aggregate resources, Alma map area (21 H/10); N.B. Dept. Nat. Res., Min. Res. Br., Open File Rept. 78-5.

---------- 1980: Granular aggregate resources, Saint-Andre (210/4), Aroostook (21 J/13), New Brunswick; N.B. Dept. Nat. Res., Min. Res. Br., Open File Rept. 80-6.

Weeks, L.J. 1957: The Apralachian region; pp. 123-205 *in* Geology and economic minerals of Canada, *ed.* C.H. Stockwell, Geol. Surv. Can. Econ. Geol. Ser. No. 1 (4th Ed.).

Welstead, J. 1971: Morphology and evolution of the Bay of Fundy with emphasis on changes of sea level during the Quaternary; Ph.D. thesis, Bristol Univ.

Williams, H., Kennedy, M.J., and Neale, E.R.W. 1972: The Appalachian structural province; pp. 181-262 *in* Variations in tectonic styles in Canada, *eds.* R.A. Price and R.J.W. Douglas, Geol. Assoc. Can., Special paper no. 2.